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EARLY DEVONIAN AND LATE SILURIAN FORMATIONS OF SOUTHEASTERN PENNSYLVANIA

BY CHARLES K. SWARTZ AND FRANK M. SWARTZ

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ABSTRACT

The area studied is part of a separate Silurian province which extends south of the coal measures from the Schuylkill River to southeastern New York. It is divided into two parts, southern and northern, each characterized by distinctive faunas and lithology.

Numerous sections are studied embracing beds from the Silurian Bloomsburg to the base of the Middle Devonian Onondaga; their formations are described, faunas given, and beds correlated.

The Middle Devonian Palmerton sandstone and Bowmanstown chert of the southern area, long designated Oriskany, are correlated with the Esopus shale of the northern area. The underlying Oriskany sandstone is present in different facies in the southern and northern areas.

The Helderberg and Keyser groups differ in facies in the two areas. They are briefly described and correlated.

The Silurian Poxono Island shale and Bossardville limestone are present in both areas. They are similar to, and homotaxial with, the Wills Creek shale and Tonoloway limestone, respectively, of central Pennsylvania and Maryland but lie in a different province, and the corresponding formations differ in their time limits. Both are believed to be later than the Wills Creek shale.

The formations above the Bloomsburg thin progressively and disappear southwestward, but the Bloomsburg is continuous throughout the area.

The Keyser limestone presents both Silurian and Devonian aspects. It is tentatively made Silurian, but the determination of its true age must await further study of its relation to the European sequence.

INTRODUCTION

SCOPE OF DISCUSSION

The authors have been engaged, for a number of years, in the investigation of the Silurian of the central Appalachian area. In pursuing this work they have been led to study the early Devonian strata also, since

here, as in most other regions, the problem of the Siluro-Devonian boundary and the relations of the Devonian to the Silurian formations are involved.

A preliminary publication (Swartz and Swartz, 1931, p. 621-662) was devoted to the early Silurian formations of southeastern Pennsylvania. The present will consider the later Silurian and early Devonian of the same area.

GEOLOGIC PROVINCE

The late Silurian and early Devonian formations are well developed and extensively exposed in central Pennsylvania. They are absent from the gap of the Susquehanna River north of Harrisburg to Swatara Gap, 23 miles farther northeast, due, probably, to unconformity. They reappear again near the Schuylkill River and increase in volume eastward to the Delaware. From the viewpoint of the late Silurian and early Devonian deposits the latter region may be said to constitute a distinct province which extends south and east of the coal measures from the Schuylkill River northeastward into southeastern New York. This province may be further divided into two subprovinces—a northern and southern—each characterized by distinctive faunas and sediments.

EARLIER INVESTIGATIONS

The most important early work was done by I. C. White, whose report, published in 1882, remains of great value today. J. B. Reeside, Jr. (1918), later made a critical investigation of some of these beds in central Pennsylvania but did not study the region under consideration, though his contribution has a valuable bearing upon it. In 1929 F. M. Swartz published a report dealing with the Helderberg group, in which he described several sections in eastern Pennsylvania. He has recently (1939, p. 29-91) presented additional information in a discussion of the Helderberg and Keyser formations of the State.

In 1937 the authors (1938, p. 117) showed that much of the supposed Oriskany of eastern Pennsylvania is actually Middle Devonian and proposed for these beds the name Palmerton sandstone and Bowmanstown chert. The conclusions presented were accepted and used by A. B. Cleaves (1939, p. 92-130) and Bradford Willard (1939, p. 131-160) in their recent and valuable studies of the Oriskany and Middle Devonian of Pennsylvania. The evidence supporting the post-Oriskany age of the Palmerton and Bowmanstown formations is presented more fully at this time, together with data on the Helderberg and Keyser beds supplementary to that recently published by F. M. Swartz, and also with much new information concerning the Oriskany, Bossardville, Poxono Island, and Bloomsburg formations.

ORDER OF DISCUSSION

The various sections will first be described, and the larger relations presented by them will then be considered.

DESCRIPTION OF SECTIONS

GENERAL STATEMENT

Insofar as possible, sections were studied sufficiently close together to permit the ready following of faunas and formations throughout the region. The sections are arranged in three groups: (1) southwestern sections at Auburn and Schuylkill Haven; (2) sections in the southern ridge and its southwestern extension; (3) sections in the northern ridge. All detailed sections by the authors are in Pennsylvania.

The senior author was assisted in many measurements by Charles D. Swartz. Identifications of Ostracoda are by F. M. Swartz.[1]

The locations of the sections are shown on the accompanying map (Fig. 1). The Roman numerals correspond with those on the columnar sections (Pl. 1).

SECTIONS IN SOUTHWESTERN AREA

Auburn, Schuylkill County (Pottsville quadrangle).—The section described is exposed along a county road on the east side of the Schuylkill River, one mile southeast of Auburn.

	Thickness, feet	
	Beds	Total
Devonian System		
Marcellus shale		
Thick-bedded black shale, *Styliola fissurella* (aa), thickness not measured.		
Concealed		74
Onondaga limestone		
Blue-gray limestone. N. 67° E., 27½° NW. Fossils numerous, including *Atrypa reticularis, Athyris spiriferoides, Leptocoelia acutiplicata, Pentagonia unisulcata, Schuchertella pandora, Spirifer varicosus, Stropheodonta (Leptostrophia) perplana, Phacops cristata*	18.5	26
Concealed	7.5	7.5
Total thickness Onondaga limestone		26
Lower Devonian		
Oriskany sandstone		
Thick-bedded coarse white conglomeratic sandstone, pebbles commonly ⅛, rarely ½ inch diameter, bearing *Rhipidomella musculosa*		3
Silurian System		
Bloomsburg red beds		
Alternating red and green sandstone and shale, N. 67½° E., 23° NW. (average three observations)		27
Red sandstone and shale; great thickness, not measured.		

[1] Measurements were made with tape and hand level, unless otherwise stated. Bearings magnetic. Work by F. M. Swartz on sections at Schuylkill Haven and on the Delaware River has been supported by the Pennsylvania Topographic and Geologic Survey.

This section is conspicuous for the absence of many formations, present elsewhere, between the Bloomsburg and Oriskany sandstones.

Schuylkill Haven, Schuylkill County (Pottsville quadrangle).—Two sections, embracing the uppermost Silurian and Lower Devonian, are

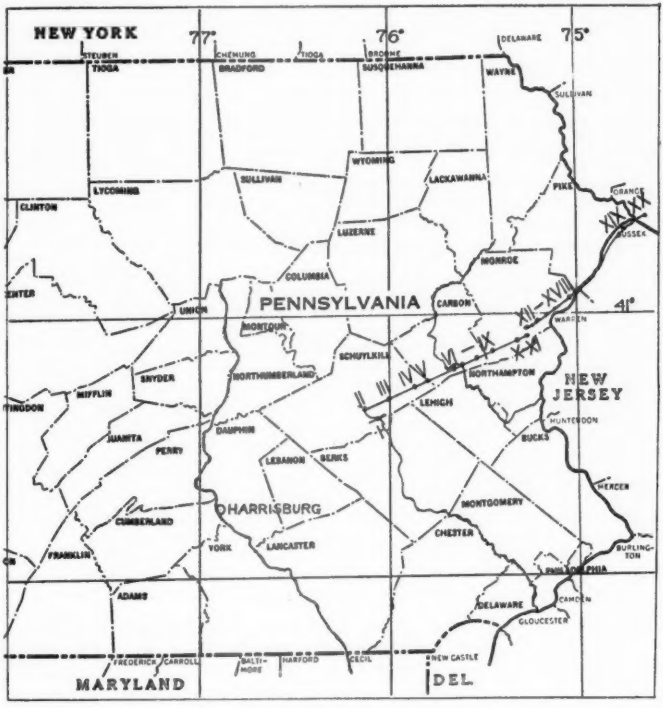

FIGURE 1.—*Map showing locations of sections*

I. Auburn; II. Schuylkill Haven; III. New Ringold; IV. Snyders; V. Andreas; VI. Bowmanstown; VII. Palmerton; VIII. Little Gap; IX. Kunkletown; X. Saylorsburg; XI. Bossardville; XII. Godfrey Ridge; XIII. Stormville; XIV. Fox Gap; XV. Delaware Water Gap; XVI. Shawnee; XVII. Poxono Island; XVIII. Decker's Ferry; XIX. Nearpass; XX. Trilobite Mountain.

exposed in the vicinity of Schuylkill Haven, the first is at a small quarry, the second in a cut of the Pennsylvania Railroad, a short distance above the quarry.

I. SECTION AT QUARRY

An excellent section, measured by F. M. Swartz, is exposed in and north of a small quarry, on the road leading from Schuylkill Haven to

Adamsdale, 1.3 miles in an airline northeast of Schuylkill Haven. The road bends to the left, in going east, at this place.

Section in quarry

Devonian System

Oriskany sandstone

	Thickness, feet	
	Beds	Total
Concealed.		
Thick-bedded gray conglomeratic sandstone, pebbles up to ¼ to ½ inch in diameter, occasionally 1 inch		5

Devonian or Silurian System

Keyser limestone

Thin to medium-bedded, rather fine-grained, green to gray sandstone and some interbedded arenaceous shale; contact with overlying sandstone very irregular	12	44
Interbedded fine-grained calcareous sandstone and arenaceous shale. At base occur *Camarotoechia lithchfieldensis?* (r), *Chonetes jerseyensis* var. (aa), *Kloedenia* sp. (rr)	4	32
Impure limestone varying to calcareous shale, weathering thin-bedded	10	28
Blue impure limestone, tending to become somewhat lumpy above. Fossils occur above base of this unit as follows: 5 feet, *Rhynchospira globosa, Dibolbina* n. sp. (fine, r-c), *Dizygopleura* sp. (c), *Halliella* sp. (rr), *Kloedenia* n. sp. (double ornamentation, c), *Kloedenella* sp. (c), *Primitiopsis? limbata* n. sp. (c), *Welleria* sp. (lobation nearly obsolete, c), *Zygobeyrichia barretti* (c); 12 feet, *Dizygopleura sp.* (c), *Kloedenia* sp. (r), *Kloedenella* sp. (a), *Primitiopsis? limbata* n. sp. (c), *Zygobeyrichia barretti* (c), *Zygobeyrichia* sp. (c); 17 feet, *Leperditia* sp. (r-c), *Dizygopleura* sp. (a), *Eukloedenella?* sp. (rr), *Kloedenia centricornis* and var. (a), *Kloedenia* sp. (c), *Kloedenella* sp. (a), *Primitiopsis? limbata* n. sp., *Welleria* sp. (r), *Zygobeyrichia barretti* (c), *Zygobeyrichia?* sp. (r); at top occur Bryozoa (c), *Beyrichia* sp. (rr), *Dizygopleura* sp. (c), *Eukloedenella?* sp. (c), *Halliella* sp. (rr), *Kloedenia centricornis* and vars. (median knob rounded, but surmounted by a slender spine) (a), *Kloedenella* sp. (c), *Leperditia scalaris* (fine, r-c), *Leperditia* sp. (rr), *Leperditiella?* sp. (rr), *Primitiopsis? limbata* n. sp. (c)	18	18
Thickness of beds referred to Keyser limestone		44

Silurian System

Bossardville limestone

Blue limestone, weathering thin-bedded, worked in quarry. 4 feet below top occur *Spirifer* sp. (small r), *Hormatoma* sp. (r), *Dizygopleura* sp. (a), *Kloedenia centricornis* vars. (c), *Primitiopsis? limbata* n. sp. (r)	25	46
Section continued along highway just north of quarry.		
Thick-bedded limestone, varying to calcareous sandstone, when fresh blue and green variegated, weathering thin-bedded. A spring issues 10 feet above its base. Fossils occur above base of this unit as follows: 3 feet, *Kloedenia centricornis* (rr); 18 feet, Linguloid brachiopod (c), *Hormatoma* sp. (c), *Dizygopleura* sp. (c), *Halliella* sp. (r), *Kloedenia centricornis* and var. (a), *Kloedenella* sp., *Leperditia* sp. (c), *Primitopsis? limbata* n. sp. (c); at top Bryozoa small (c), *Hormatoma* sp. (r-c), *Halliella* sp. (r), *Kloedenia centricornis* and vars. (a), *Primitiopsis? limbata* n. sp. (c-a)	21	21
Thickness of beds referred to Bossardville limestone		46

Bloomsburg red beds	Thickness, feet Beds	Total
Concealed, 2 feet thick-bedded green sandstone, 5.5 to 7.5 feet above base	18	24
Hackly red siltstone, 1.5 feet thick at top, concealed below	6	6
Great thickness of red sandstone and arenaceous red shale is exposed along road westward. Not measured.		

II. Section on Railroad

An uninterrupted section, extending from the upper part of the Bloomsburg to the lower beds of the Keyser, is seen in the cut of the Pennsylvania Railroad a short distance above the quarry, supplementing the section already described. The measurements are by C. K. Swartz, assisted by C. D. Swartz. Bearings magnetic 1938.

Devonian System

Oriskany sandstone

Concealed. Thick-bedded gray conglomeratic sandstone,[2] containing indistinct fossils. A. B. Cleaves (personal communication) reports *Spirifer arenosus* at this horizon.

Devonian or Silurian System

Keyser limestone	Thickness, feet Beds	Total
Concealed. Much calcareous shale loose in talus, in lower part, weathering into greenish fragments and bearing numerous Ostracoda	40	44
Impure greenish, somewhat lumpy limestone; 1 foot above base occur Bryozoa (a); 3 feet above base *Camarotoechia lamellata, C. litchfieldensis, Rhynchospire globosa, Megambonia*? sp.	4	4
Thickness of beds referred to Keyser limestone		44

Silurian System

Bossardville limestone		
Blue limestone weathering thin-bedded, some beds oölitic, upper 2 feet somewhat lumpy, some interbedded calcareous shale. *Leperditia* sp. (a), N. 79° E., 27° SE. (average three observations)	35	44.5
Blue limestone, laminated, in part oölitic	2.5	9.5
Calcareous shale	2.5	7
Bluish-gray, thin-bedded, laminated limestone, in part oölitic	4.5	4.5
Thickness of beds referred to Bossardville limestone		44.5

Bloomsburg red beds		
Greenish calcareous mud rock, hackly fracture, weathering thin-bedded; 1-foot sandstone from 6.5 to 7.5 feet below top, greenish arenaceous shale at base		15
Red sandstone and shale, some greenish beds, seen in railroad cut west of measured section, thickness not measured.		

[2] This bed is located as follows: traversing from the top of the *Rhynchospira globosa* zone of the underlying units, S.55°E.210 feet, along the railroad tracks; thence N.19°E.80 feet, rise 11.5 feet to the approximate outcrop of the bed. The sandstone is exposed in place in a small prospect a short distance farther northeast.

The 15 feet of greenish beds, here made Bloomsburg, might be referred to the Poxono Island shale. They are similar, however, to lower green zones within the red beds.

The limits here assigned to the Bossardville and Keyser are uncertain, and the identification of the lower limestone beds as Bossardville is insecure.

The reference of the latter to the Bossardville is based upon (1) the lithology, (2) the occurrence in them of *Hormatoma* sp., a genus characteristic of the Tonoloway elsewhere, and (3) the connection of these beds by rather close sections with the typical Bossardville limestone eastward. Against their reference to the Bossardville is (1) the occurrence in them of Ostracoda of known Keyser, or Cobleskill, species, or at least of very close variants, especially *Kloedenia centricornis* and *Leperditia scalaris*. (2) These species and the new *Primitiopsis ? limbata*, and probably others, also, range upward into the Keyser beds of this locality. All the beds here described as Bossardville may hence be Keyser. The answer to these questions can be found only by further study of the zonal relations of the ostracod species here cited.[3]

SECTIONS IN SOUTHERN RIDGE AND ITS SOUTHWESTERN EXTENSION

New Ringold, Schuylkill County.—Two sections are exposed in the vicinity of New Ringold, one ½ mile southwest of New Ringold, in the Pottsville quadrangle, the other 2¼ miles east of the village in the Hamburg quadrangle.

I. Section ½ Mile Southwest of New Ringold

The section is seen along the county road which leaves the State highway ½ mile west of New Ringold and runs thence southward, on the west side of the Little Schuylkill River, toward Rauschs. The section begins about 1800 feet south of the intersection of the county road with the State highway from New Ringold to Orwigsburg, and 250 feet north of a marked bend of the road to the west. The strata are concealed north of this point. The section continues southward 3600 feet to the top of the Clinton formation.

This section is of unusual interest since it gives the best measurement of the thickness of the Bloomsburg red beds observed in this area, good exposures of the limits of that formation being rare. For this reason the measurements of the thickness of the Bloomsburg are given in detail.[4]

[3] In the case of *Kloedenia centricornis* it may be possible to distinguish the rather slender-spined form, here referred to that species, from the stouter-spined variant, which occurs in the higher part of the Keyser limestone of other areas.

[4] Bearings magnetic 1928. Traverse directions: 0-250 feet S. 43° E., 250-400 feet S. 10° E., 400-750 feet S. 15° W., 750-1050 feet S. 2° W., 1050-1450 feet S. 2° E., house west of 1200 feet, 1450-1900 feet S. 21° E., center cement bridge 1731 feet, 1900-2550 feet S. 11° E., drain 2515 feet, 2550-3000 feet S. 37° E., top grade 2977 feet, 3000-3600 feet S. 49° E.

Devonian System Oriskany (?) sandstone	Thickness, feet Beds	 Total
Concealed. Many boulders of massive gray conglomeratic sandstone lie in field on west. No fossils seen. Base at 50 feet traverse....	50	50
Devonian or Silurian System Helderburg? and Keyser beds		
Concealed. Fragments of soft, buff, calcareous shale in field on west, age unknown. Base at 400 feet traverse. 100 to 125 feet N. 8° W. from top of the Bloomsburg red beds and about 40 feet stratigraphically above the base of this unit occur, loose, in field, *Dizygopleura* sp. (aa), *Kloedenella* sp. (c), *Kloedenia centricornis* vars. (a)	225	225
Silurian System Bossardville limestone		
Concealed on road. Thin-bedded blue limestone and calcareous shale are exposed in small quarry west of road. Base unit at 444 feet traverse on road	36	36
Bloomsburg red beds		
Variegated green and red sandstone and shale, N. 86° W., 55° NE., N. 81° W., N. 82° W. (average N. 83° W., 55° NE.) Base at 502 feet traverse	47	1815
Red shale and sandstone, concealed in part. Base at 3200 feet traverse. Average strike and dip used, N. 84° W., 48° NE. (average five strikes, 12 dips).....	1768	1768
Total Thickness Bloomsburg red beds.....		1815
Clinton formation		
Fragments gray sandstone loose on bank, gray soil in pit at base, much red soil in unit. Base at 3483 feet traverse.....	117	152
Concealed above, variegated red sandstone 3535 to 3600 feet at end of traverse	45	45
Thickness Clinton formation described.....		152

Concealed.

Loose boulders of conglomeratic sandstone extend for several hundred feet on the west side of the road, north of the point of beginning. No fossils were seen in them and their age is problematical. Their stratigraphic relations suggest the Oriskany, but they may represent either the Oriskany or Palmerton, or both.

North of the sandstone boulders are loose pieces of white chert bearing a New Scotland fauna, including: *Eospirifer macropleura, Meristella arcuata* var. very large and convex (a), *Schuchertella woolworthana* (a), *Spirifer perlamellosus, Stropheodonta (Leptostrophia) beckii, Uncinulus vellicatus?*, and *Platyceras* sp.

II. Section in Quarry 2¼ Miles East of New Ringold

An excellent section of Bossardville and Keyser beds is exposed in the eastern of two small limestone quarries, 2¼ miles east of New Rin-

gold and 0.4 mile south of the highway to Palmerton. The section described is seen at the west end of the quarry. The beds strike due east and west (magnetic 1935) and dip 57° N. The exposure is supplemented in fields near a dwelling ½ mile west of the quarry. Measurements are by C. K. Swartz, assisted by C. D. Swartz.

Section in field west of dwelling, ½ mile west of quarry

	Thickness, feet	
	Beds	Total
Devonian System		
Oriskany(?) sandstone		
Concealed. Numerous boulders of massive gray conglomeratic sandstone loose in field. No fossils observed.		
Age unknown		
Concealed. Fragments of reddish sandstone loose in field toward top. Horizontal distance from approximate base of overlying conglomeratic sandstone to top of Bloomsburg red beds, measured perpendicular to strike, about 300 feet. Thickness of interval, using dip 57° N., is about 250 feet, making thickness of beds between top of Keyser beds exposed in quarry and base of the conglomeratic sandstone approximately	175	175
Section in quarry		
Devonian or Silurian System		
Keyser limestone		
Calcareous shale, weathering yellowish, some interbedded layers of rotten very argillaceous limestone carrying at base *Rhynchospira globosa* var., *Actinopteria emacerata;* 1 foot above base, *Chonetes jerseyensis, Rhynchospira globosa* var. (c), *Actinopteria emacerata;* 3 feet above base Bryozoa (c); 5 feet above base Bryozoa, *Rhynchospira globosa* var. (c). Undetermined Ostracoda are profuse in these beds	15	27.5
Concealed	12.5	12.5
Thickness Keyser beds described		27.5
Silurian System		
Bossardville limestone		
Blue limestone and some calcareous shale, lower 1.5 feet thicker-bedded, upper foot loose, Bryozoa, (c) at base	7.5	50.5
Gray limestone in a single course, calcite streaks	1	43
Thin-bedded gray limestone and calcareous shale, concealed 1 to 4 feet above base, 6 inches below top occur Bryozoa (c)	6	42
Calcareous shale, with bands of blue to gray limestone, some wavy mud cracks in upper part. Upper shales weather greenish. *Leperditia* sp. (aa), and Ostracoda, 2 to 4 feet above base	4	36
Shaly limestone; 3 inches of blue limestone at top	1	32
Dark-blue limestone; upper 2½ feet thinner-bedded, with some shale; lower 2 feet thick-bedded, with calcite streaks	4.5	31
Calcareous shale, seen in east end of quarry	3	26.5

	Thickness, feet Beds	Total
Dark-gray, calcareous shale, seen in recess in footwall. The top of of this unit forms footwall of quarry and bears many mud cracks. *Leperditia* sp. (a)	3.5	23.5
Concealed 25 feet horizontally, in field south of quarry. Soil with many greenish-gray fragments. A little intermingled red soil, perhaps from plowing. Age unknown. May be Poxono Island shale..	20	20
Total thickness of Bossardville limestone		50.5

Bloomsburg red beds

Soil filled with fragments of red shale. Great thickness, not measured.

The beds here described as Bossardville limestone consist of two parts: a lower calcareous shale and concealed, 26½ feet thick, and an upper blue limestone with subordinate amounts of calcareous shale, 24 feet thick. The lower shale might be regarded as Poxono Island shale because of its argillaceous character. The total thickness of 50 feet may be compared with 44 feet, measured at Schuylkill Haven, and 36 feet ½ mile southwest of New Ringold.

The thickness from the top of the Bloomsburg red beds to the base of the Oriskany? conglomeratic sandstone is about 250 feet in Section II. The thickness from the top of the Bossardville limestone to the base of the sandstone is about 200 feet and is in close agreement with the thickness of this interval farther east.

Snyders, Schuylkill County (Hamburg quadrangle).—Quarries were formerly worked at two localities east of the village, (1) on the county road, (2) on Pennsylvania State Highway 29.

I. Section on County Road, 1¼ Miles East of Snyders

Quarries were formerly worked on the crest of the hill 1 to 1½ miles due east of Snyders, extending nearly ½ mile in a line NE.-SW., some of the workings being shown on the U. S. topographic map. The line joining them crosses the road about 90 feet southeast of the top of the grade on the road. Three hundred sixty feet northeast of the intersection of the road and quarries, the quarry trench strikes N. 82° E. Nineteen feet of shaly limestone are exposed in the quarry at this point, dipping 90 degrees and abounding in *Leperditia* sp. Six hundred feet southwest of the intersection the trench strikes N. 87° E., and the beds dip, apparently, 60° NW. The general strike of the quarry trench, to a point 600 feet southwest of the intersection with the road, is N. 82° E.

The traverse begins at the outcrop of the New Scotland white chert, on the county road, 680 feet northwest of the intersection of the center line of the quarry and the county road. It extends thence 232 feet S. 78° E.

along the road, then 145 feet N. 26° W. along a lane. A side traverse was run from station 0 + 85 feet on the road, 455 feet S. 3° E. across fields, to the top of the exposure of the Bloomsburg red beds.[5]

The section is described as follows:

Devonian System

Section along lane

Oriskany(?) sandstone and New Scotland chert	Thickness, feet Beds	Total
Numerous boulders of massive gray conglomeratic sandstone, loose, in upper part. Some loose white chert. Lower part concealed. Fields smooth to the northwest. 317 to 377 feet traverse. The base of this unit is at the approximate horizon of zero feet traverse on road		50
Section along road		
Much white chert loose in road at beginning of traverse, bearing a New Scotland fauna including *Coelospira concava* (a), *Meristella* n. sp. 1 and 2, *Uncinulus vellicatus* var. (a). Concealed below		
Thickness of beds from exposure of New Scotland chert to top of loose Oriskany(?) sandstone		50
Age undetermined		
Concealed. 0 feet to 85 feet traverse	25	220
Section across field from station 0 + 85 feet		
Concealed to north side of quarry trench, at 310 feet traverse	195	195
Total thickness beds from top Bossardville limestone, quarried, to New Scotland chert, approximately		220

Silurian System

Bossardville limestone

Concealed across quarry trench, distance estimated. To 360 feet traverse	43	43
Poxono Island shale		
Concealed, soil light-colored. (Traverse farther southeast). To 540 feet traverse	156	156
Bloomsburg red beds		
Wide band of red soil of Bloomsburg type, farther south, with lighter bands of Wills Creek type. Great thickness not measured.		

II. Section of Pennsylvania State Highway Number 29

Pennsylvania State Highway 29, leading from Tamaqua to Allentown, crosses the Pottsville branch of the Lehigh Valley Railroad 800 feet northeast of West Penn Station, measured along the railroad tracks. Southward it passes through the letter P of the name West Penn on the U.S. topographic map.

A limestone quarry, now abandoned, was formerly worked on the south side of the State Highway 950 feet (auto speedometer) southeast

[5] Measured by C. K. Swartz, assisted by C. D. Swartz. Bearings magnetic 1939. Strike and dip used in calculating thickness N. 82° E., 60° NW.

of the railroad crossing. A county road crosses the highway 180 feet southeast of the center of the quarry.

The measured traverse begins at the point of intersection of a line along the west wall of the quarry, produced, with the south side of the highway and extends thence 300 feet N. 57° W. along the south side of the highway. Beyond this point the section is concealed. Strike and dip in quarry, N. 74° E., 90°, are used in calculation. The measurements are by C. K. Swartz and C. D. Swartz. Bearings magnetic 1939.

	Thickness, feet	
	Beds	Total
Devonian System		
New Scotland		
Concealed. Numerous fragments of white chert, loose on north side of road on strike from station 300 feet of traverse, contain *Meristella arcuata, Meristella* sp., *Eospirifer macropleura, Spirifer perlamellosus?, Uncinulus abruptus* (a).		
Age undetermined		
Concealed across lane. To 300 feet traverse	19	226
Concealed. Numerous fragments of red-brown sandstone in talus, especially in lower part of unit; 1 foot of red-brown conglomeratic sandstone, loose, at base; a few fragments of gray sandstone from 236 feet traverse to top; some pieces of gray conglomerate in upper 27 feet. To 275 feet traverse	127	207
Devonian or Silurian System		
Keyser formation		
Red-brown thin-bedded sandstone and shale. The Andreas red beds. To 106 feet traverse	28	80
Greenish arenaceous shale and thin-bedded calcareous sandstone, 2 feet greenish calcareous sandstone at top. To 69 feet traverse	6.5	52
Very arenaceous mudstone, weathering greenish, *Leperdita* sp. (aa) in lower foot. To 60.5 feet traverse	9.5	45.5
Calcareous shale and some impure arenaceous gray limestone, bearing, above base, ¾ feet (23 feet traverse), *Dizygopleura* sp. (c-a), *Kloedenia* sp. (r), *Zygobeyrichia barretti* (c), *Zygobeyrichia?* sp. (near *Kyammodes swartzi*); 2¼ to 3 feet (25 to 26 feet traverse), Bryozoa, *Rhynochospira globosa, Uncinulus* sp., *Megambonia* sp. (r), *Leperditia* sp., Ostracoda; 4½ feet (28 to 29 feet traverse), Bryozoa, *Rhynchospira globosa* (c), *Megambonia* sp. 2, *Leperditia* sp. To 48 feet traverse	19.5	36
Calcareous shale and some beds lumpy limestone, partially concealed, bearing, above base, 5 feet, Ostracoda; 8½ feet, Bryozoa, Ostracoda; 12 feet (20 feet traverse), *Dizygopleura* sp. (c), *Kloedenia* sp. smooth, (c), *Kloedenia* sp. aff. *K. centricornis*, but median knob rounded, *Kloedenella* sp. (aa), *Zygobeyrichia barretti* (a), *Zygobeyrichia?* sp. (near *Kyammodes swartzi*) (a), *Leperditia* sp.; 12½ feet (20.5 feet traverse), Bryozoa, *Rhynchospira globosa, Uncinulus* sp.; 13 feet (21 feet traverse), *Leperditia* sp. (aa). To 22 feet traverse	13.5	16.5
Heavy-bedded lumpy limestone, forming northwest wall of quarry N. 74° E., 90°. To 3 feet traverse	3	3
Thickness Keyser formation described		80

Silurian System

Bossardville limestone

Section in quarry	Thickness, feet: Beds	Total
Quarry rock concealed in large part. Some compact dark-blue limestone seen in lower part. Upper 10 feet of this unit may be Keyser	35	43
Shaly gray limestone	8	8
Thickness quarry limestone measured		43

The section is concealed for 600 feet, southeastward from the quarry, along the highway, beyond which are red shales and sandstone of Bloomsburg type, interbedded with some lighter-colored shale of Wills Creek type.[6]

Andreas, Schuylkill County (*Hamburg quadrangle*).—Three sections have been studied in the vicinity.

I. Quarry of J. Rudolph

The quarry of J. Rudolph is situated ½ mile, in an airline, south of Andreas. The quarry is located west of the county road, near the axis of a small anticline. It affords the best section of the Keyser and Bossardville limestones observed by the authors between the Schuylkill and Lehigh Rivers. It exposes the following section:

Devonian System	Thickness, feet: Beds	Total
Concealed. Boulders of thick-bedded gray conglomeratic sandstone cover the hillside. No fossils were observed. Presumably the Palmerton sandstone, though some may be derived from the Oriskany. This sandstone is opened in a small quarry along the road south of the limestone quarry.		
Devonian or Silurian System		
Keyser formation		
Concealed. Numerous fragments of red sandstone and shale lie loose on hillside above the quarry, associated with boulders from the higher conglomeratic gray sandstone. These beds extend some 70 feet above the underlying unit. Their Keyser age is uncertain. The Andreas red beds.		
Greenish-gray calcareous sandstone, upper part weathering thin-bedded; seen above quarry on its west side. The lower 3 feet of this sandstone is also exposed near top of quarry workings, being thin-bedded, with partings of calcareous shale. Some imperfectly preserved fossils include Bryozoa, *Chonetes jerseyensis*, *Strophеodonta* (*Leptostrophia*) *bipartita*	9	70
Calcareous shale and shaly limestone	2	63

[6] The measurement of the thickness of the beds between the Keyser and New Scotland cherts is open to question. The thickness given is much greater than in neighboring sections; the New Scotland cherts are loose, and intervening beds concealed. Against this is the agreement in measurement in the two sections here given.

	Thickness, feet Beds	Total
Thick-bedded dense dark-blue argillaceous limestone, weathering lumpy, breaking generally with oblique schistosity. These beds become increasingly arenaceous, above, weathering greenish; 37 feet above base occur Bryozoa, *Chonetes jerseyensis, Rhynchospira globosa;* 32 to 33 feet, fossil cavities (aa); 22 feet, Ostracoda (aa), *Rhynchospira globosa;* 15 feet, Ostracoda loose (aa)	55	61
Gray somewhat irregularly bedded limestone, calcite streaks in places, Bryozoa. This bed may be of Bossardville age	4	4
Total thickness Keyser formation, exclusive of Andreas red beds of uncertain age		70
Silurian System		
Bossardville limestone		
Strongly laminated blue limestone, weathering into prismatic pillars 5 to 7 feet above base	10	100
Dense dark, blue-gray limestone, many calcite streaks	52	90
Argillaceous gray limestone with oblique schistosity, lower 2 feet and upper 6½ feet weather yellowish	14.5	38
Gray limestone band	0.5	23.5
Gray limestone, increasingly impure above, where it weathers with yellowish lines	7	23
Dark-gray limestone, calcite streaks in lower part	12	16
Argillaceous dark-gray limestone, lower 2 feet weathering yellowish, upper 2 feet with oblique schistosity	4	14
Total thickness Bossardville limestone		100
Concealed to quarry floor		6

The Bossardville limestone as described above consists of two rather clearly defined parts, an upper dense and purer limestone, 62 feet thick, and a lower more argillaceous limestone, 38 feet thick.

II. Quarry of William Loch

This quarry is situated ½ mile southeast of Andreas, on the east side of a small stream, and about ¼-mile northeast of the quarry of J. Rudolph. It exposes essentially the same section as that already described but adds interesting details.

The following section was measured at the north end of the quarry.

Devonian or Silurian System Keyser formation	Thickness, feet Beds	Total
Red sandstone forming ledges on hill just above top of quarry face, concealed in part. N. 49° E., 20½° SE. (average two observations). The age of these beds is uncertain. The Andreas red beds (type locality)	17.5	81
Concealed	23.5	67.5
Calcareous sandstone, weathering greenish and forming a projecting ledge just above quarry face. Thickness estimated	8	44

	Thickness, feet	
	Beds	Total
Calcareous shale, some thin bands of sandstone, many solution cavities in lower foot, a 6-inch bed of sandstone at base. Farther south in quarry a highly fossiliferous band occurs 1½ feet above base where it carries Bryozoa, *Camarotoechia* sp., *Chonetes jerseyensis, Rhynchospira globosa, Rhynchospira?* sp., *Spirifer* sp., *Stropheodonta (Leptostrophia) bipartita*	18.5	36
Strongly laminated calcareous shale with two 3-inch bands of limestone, 12 and 6 inches, respectively, above its base	8	17.5
Argillaceous limestone to calcareous shale, with many 1-inch bands of purer, nodular limestone which weathers into holes	8	9.5
Very lumpy gray limestone, 1 foot above its base occur *Rhynchospira* sp., *Dizygopleura* sp. (a), *Kloedenia* sp. (like *K. centricornis*, but round knob) (a),, *Kloedenia* sp. smooth (a), *Kloedenella* sp. smooth, with crest (aa), *Zygobeyrichia* sp. (near *Kyammodes swartzi*)	2.5	6.5
Dark-blue-gray limestone, its lower 4 inches oölitic, middle thin-bedded, 6 inches shale at top. Limestone contains Bryozoa (a) and *Leperditia* sp. (a); 2 feet above its base occur *Kloedenella* sp. (a) poor, *Zygobeyrichia barretti* (r)	4	4
Thickness Keyser limestone, exclusive of red beds of uncertain age		67.5
Silurian System		
Bossardville limestone		
Strongly laminated blue-gray limestone, lower 1½ feet weathering into vertical, prismlike structures	5	100
Calcareous shale with oblique schistosity	4	95
Laminated blue-gray limestone, much concealed below, in cave	24	91
Concealed to base of section exposed in quarry of J. Rudolph	67	67
Approximate thickness of Bossardville limestone		100

At the south end of the quarry, near the top of the exposure of Keyser limestone, occur Bryozoa, *Camarotoechia litchfieldensis, Camarotoechia* sp., *Chonetes jerseyensis?, Rhynchospira globosa, Spirifer* sp., *Stropheodonta (Leptostrophia) bipartita.*

The name Andreas red beds is proposed, provisionally, for the red sandstone forming ledges above the quarry at this place. It is known only at Snyders and Andreas but is so definite a body that it seems desirable to give it a separate name. Its age is unknown, but it overlies beds bearing the Keyser fauna and is overlain by loose fragments of chert bearing a New Scotland fauna in the road north of the quarry and at Snyders. It is placed provisionally in the Keyser formation.

III. Section on Farm of C. A. Gombert

The farm of C. A. Gombert is located 0.9 miles southwest of Andreas, on a road leading westward to Millers. The dwelling is on Devonian black shale near the axis of a sharp syncline. Going southward the

Onondaga "paint-ore" horizon is seen, underlain by thick-bedded, gray, conglomeratic Palmerton sandstone, beneath which white cherts crop out.

The cherts bear a large New Scotland fauna including *Mariacrinus* cf. *stoloniferus, Coelospira concava* (aa), *Dalmanella* sp., *Eatonia medialis, Leptaena rhomboidalis, Meristella arcuata* (aa), *Meristella* n.sp.1(c), *Meristella* n.sp.2, *Schuchertella radiata?, Schuchertella woolworthana* (a), *Eospirifer macropleura, Spirifer cyclopterus?,* numerous small specimens, young?, *Spirifer perlamellosus, Spirifer* sp., *Strophonella punctulifera, Uncinulus abruptus, Uncinulus vellicatus* var.(aa), *Dalmanites* sp., and *Phacops logani.*

The conglomeratic sandstone and the chert occur as loose pieces in the soil but litter the surface in such profusion in well-marked zones, that it is clear all are near their outcrops. The Devonian black shale is nearly vertical. The distance perpendicular to the strike from the top of the conglomeratic sandstone to the fossiliferous chert is about 175-200 feet. The cherts lie hence approximately 175-200 feet stratigraphically below the top of the Palmerton sandstone. The cherts are of New Scotland age.

A similar rich New Scotland fauna was collected in white cherts lying loose above the Andreas red beds, on the road ascending the hill to the east, a short distance north of the William Loch quarry. The following species were found there in cherts: cyathophylloid corals(aa), *Edriocrinus pocilliformis* (aa), *Coelospira concava* (aa), *Dalmanella* sp., *Eatonia singularis, Leptaena rhomboidalis* (c), *Meristella arcuata* (a), *Meristella* n.sp.1(c), *Meristella* n.sp.2, *Nucleospira ventricosa, Rensselaeria cf. subglobosa, Rhynchospira formosa?, Schuchertella* sp., *Spirifer cyclopterus* (aa), *Spirifer perlamellosus* (c), *Spirifer* sp., *Strophonella punctulifera, Uncinulus abruptus, Uncinulus vellicatus* var. (aa).

This fauna, like that collected in the cherts at Gombert's, is clearly of New Scotland age. Certain species, however, suggest late New Scotland to Becraft.

Bowmanstown, Carbon County (Mauch Chunk quadrangle).—Two sections are exposed ¾ mile southeast of Bowmanstown, one on the State road leading to Palmerton, the other in the cut of the C.R.R.N.J. just south of the highway.

I. Section on State Highway Leading from Bowmanstown to Palmerton

The section begins ½ mile southeast of Bowmanstown and extends thence eastward along the highway. A sharp syncline repeats the section farther north.

Devonian System	Thickness, feet	
Marcellus shale and Onondaga limestone	Beds	Total
Black shale, thickness not measured.		
Concealed, ore prospects 28 and 48 feet above base	73	73
Palmerton sandstone		
Massive coarse gray iron-stained sandstone, its top forming a high cliff north of road. Much of the sandstone is soft and friable; lower beds partially concealed		125
Bowmanstown chert		
Interbedded coarse gray sandstone and white chert	5	67
White chert, lowest beds greenish, sandy; lower 10 feet richly fossiliferous, bearing at base *Spirifer macrus* var., *Tentaculitis scalariformis;* 3 feet above base *Spirifer macrus* var.	14	62
Greenish argillaceous sandstone and much black chert, thickness about	40	48
Concealed	8	8
Approximate thickness Bowmanstown chert		67
Oriskany sandstone		
Massive gray conglomerate, pebbles up to ½ inch diameter; many imperfect fossil imprints. Well-preserved *Spirifer arenosus, Meristella rostellata* and *Meristella* sp. were obtained by extensive breakage of rocks	15	15
Thickness Oriskany sandstone exposed		15
Concealed.		

II. Section on Central Railroad of New Jersey

This section is seen in the railroad cuts immediately south of the section on the highway and is supplementary to that section. The traverse begins at a drain pipe beneath the highway, just west of the Romney-Palmerton contact, ½ mile southeast of Bowmanstown, and extends southeastward along the north rail of the tracks except as noted. The strike is taken as N. 64½° E. from the topographic map. The dip varies from 90° to 40° NW. The section was measured by C. K. Swartz, assisted by C. D. Swartz. Traverse measurements are to bases of units.

Devonian System	Thickness, feet	
	Beds	Total
Concealed. A prospect on highway at 16 feet traverse. Stone retaining wall of railroad begins at 26 feet traverse. To 44 feet traverse.		
Palmerton sandstone		
Massive coarse gray disintegrating sandstone with pink hematitic stains in upper part. Upper 52 feet (to 105 feet traverse) concealed along railroad by a retaining wall. To 191 feet traverse	126	126

	Thickness, feet	
	Beds	Total
Bowmanstown chert		
Gray sandstone, some chert courses. To 196 feet traverse	4	61
Thick-bedded gray shale and light-colored chert, pink-stained near top, dip lower part 40° N.; concealed along track, from 230 to 252 feet traverse, by wall. To 265 feet traverse	47	57
Concealed to 277 feet traverse	10	10
Concealed, traverse along strike SW., to 303 feet.		
Total thickness Bowmanstown chert		61
Oriskany sandstone		
Massive gray coarsely conglomeratic sandstone, concealed in part; bearing *Meristella* sp., suggests *M. symmetrica*. Traverse along north rail on long tangent (S. 63½° E., by map)	10	10
Thickness Oriskany sandstone exposed		10
Formations from lower part of Oriskany to upper part of Bossardville limestone, inclusive, concealed. Railroad signal opposite 432 feet traverse. To 619 feet traverse	211	211
Silurian System		
Bossardville limestone		
Gray crinkly shaly limestone and calcareous shale. *Leperditia* sp. To 630 feet traverse	8	44
Impure shaly limestone and calcareous shale, buff color. To 637 feet traverse	5	36
Greenish-gray calcareous shale; 1.5 feet banded gray, oölitic limestone and calcareous shale 19.5 to 21 feet above base (646 to 648 feet traverse). To 675 feet traverse	27.5	31
Banded gray oölitic limestone and light-colored shaly limestone and calcareous shale, *Leperditia* sp. To 680 feet traverse	3.5	3.5
Thickness Bossardville limestone described		44
Poxono Island shale		
Calcareous shale, greenish to gray, or at places yellowish, concealed in large part. A pink, variegated bed 62 feet below top (765 feet traverse). Sandy yellowish shale, loose, 84 to 97 feet below top (812 to 836 feet traverse). To 963 feet traverse	206	212
Red shale. To 971 feet traverse	6	6
Concealed. Gray to greenish shale flakes in talus in upper part. Railroad signal at 1044 feet traverse. A retaining wall begins at 1078 feet, concealing beds eastward.		
Thickness Poxono Island shale described		212

Hazard, Carbon County (Mauch Chunk quadrangle).—A tunnel, formerly known as Klines' Ore Tunnel, was driven to secure paint ore at Hazard, 1½ miles west of Palmerton, Pennsylvania. The tunnel is opposite the entrance to the works of the New Jersey Zinc Company. It begins in loose drift, enters the Poxono Island shale, and terminates in the Onondaga limestone, beneath which the paint ore is

found. The tunnel is of unusual interest because it affords the most complete section known in eastern Pennsylvania of the beds from the Poxono Island shale to the Onondaga limestone. Some beds are, however, obscured by timbering.[7]

The tunnel is driven nearly at right angles to the strike. The traverse was measured with tape, beginning at the mouth (timbering) of the tunnel. The dip of the rocks toward the mouth of the tunnel is about 60° N., becoming 80° to 82° N. toward the inner face. Dips used were 60° N. from 490 to 826 feet, 72° N. from 826 to 1217 feet.

The section in the tunnel has been measured independently by the late T. A. Y. Hodgson, of Hazard, who kindly allowed the authors access to his records, and by B. L. Miller (1911, p. 53). The section of the authors is as follows:

Section in tunnel of the Prince Manufacturing Company at Hazard

Devonian System Onondaga limestone	Thickness, feet Beds	 Total
Hydraulic limestone, thickness reported in a neighboring tunnel of same company (Hill, 1887, p. 1376)	45	53
Paint ore	2–3	8
Clay. Willard (1936, p. 591) suggests that the clay is Esopus	5–6	5
Thickness of beds referred to the Onondaga		53
Palmerton sandstone		
Massive gray sandstone tending to disintegrate, some pebbles; partially concealed. Strike about N. 72° E., dip 52° NW. (average three observations). Top at 1347 feet traverse in present tunnel, elev. 567 feet. This point is about 138 feet N. 51° E. in a straight line from station 12+17 feet.[8]	35	140
Massive gray sandstone, some pebbles, dip 72° NW. (average four observations). To 1217 feet traverse, elev. 560 feet. Present tunnel turns to northeast at 12+17 feet	105	105
Total thickness Palmerton sandstone		140
Bowmanstown chert		
Arenaceous shale, sandstone, and much chert; some of chert dark. Hodgson reports 6 feet of black clay, 4 feet of hematite ore, clay, and chert at the base of these beds. To 1107 feet traverse	60	60
Oriskany sandstone		
Massive gray sandstone, upper 8 feet coarsely conglomeratic. The conglomeratic beds contain *Spirifer arenosus* and other Oriskany fossils where they are exposed along the highway and railroad tracks east of Bowmanstown. To 1047 feet traverse	25	25

[7] The authors are greatly indebted to Mr. H. E. Sanders, President of the Prince Manufacturing Company for his co-operation in the investigation, and to Mr. Wesley N. Boyer, Mining Engineer, for his assistance in measuring the thickness of the Palmerton sandstone.

[8] The original tunnel (now inaccessible) was continued from station 12+17 feet, nearly 90° to strike, to the ore body. Hodgson reports the top of the sandstone at 1260 feet traverse in it. The thickness here given is based upon his figure.

	Thickness, feet	
	Beds	Total
Helderberg Group		
New Scotland chert		
Interbedded white chert and reddish to yellow clay. To 1021 feet traverse	41	41
Coeymans (?) formation		
Thick-bedded coarse-grained friable sandstone. To 978 feet traverse	3	52
Pink and yellowish clay, some thin sandstones, much concealed. To 974.5 feet traverse	35	49
Medium-bedded fine-grained sandstone, weathering soft and yellowish to pink. Pipe (spring?) at 926 feet traverse. To 938 feet traverse	14	14
Thickness of beds questionably referred to the Coeymans formation		52
Devonian or Silurian System		
Keyser formation		
Arenaceous shale and blocky sandstone; the shale, in part, stained or colored deep red, especially at bottom and top. Three beds of hard blocky medium-bedded sandstone at 2 to 5, 18 to 20, and 25 to 27½ feet above base, respectively. The reference of these beds to the Keyser is very uncertain. The red beds may correspond to the red sandstone at a comparable position at Andreas. Unfortunately, the tunnel beds have been much affected by meteoric water, so that the red color may be due to weathering. To 923 feet traverse.[9]	29	93
Medium-bedded fine-grained yellowish to greenish sandstone, and some interbedded shale. To 893 feet traverse	5	64
Irregularly bedded arenaceous shale, weathering yellowish to greenish. To 888 feet traverse	19	59
Sandstone; "hackle teeth" on top. To 868 feet traverse	1	40
Impure lumpy limestone, weathering porous, greenish to yellowish. Bryozoa (a) on west side, 838 to 839 feet traverse, 11½ to 12½ feet above base	39	39
Thickness of beds referred to Keyser formation		93
Silurian System		
Bossardville limestone		
Hard blue laminated limestone. To 826 feet traverse	6	76
Much concealed; in part consisting of calcareous shale and blue limestone. This unit is so much concealed by timbering that it was not possible to secure measurements good enough to separate it from the underlying shale. Hodgson reports 22.6 feet traverse through his "quarry limestone." To 820 feet traverse	70	70
Thickness of beds referred to Bossardville limestone		76

[9] Both Hodgson and B. L. Miller report that these and the next overlying beds consist largely of clay and flint. Their descriptions are as follows, given in descending order from the base of the Oriskany, with traverse distances in feet: Hodgson: flint 9, clay 11, flint 23, sandstone 2, clay 24, flint 9, sandstone 4, flint 21, shale and clay 14.3, flint 6.3; total 123 feet 8 inches. B. L. Miller: clay and flint 72, sandstone 2, clay and flint 30, sandstone 4, flint 20; total 128 feet. Our traverse distance, 128 feet.

Poxono Island shale	Thickness, feet Beds	Total
Interbedded calcareous shale and some limestone; upper limit uncertain	86	215
Greenish shale. To 640 feet traverse	27	129
Red shale at top; concealed below. To 609 feet traverse	24	102
Argillaceous sandstone. To 581 feet traverse	4.5	78
Greenish shale. To 576 feet traverse	65	73.5
Red shale. To 501 feet traverse	8.5	8.5
Thickness Poxono Island shale described		215
Greenish shale; schistose cleavage; dip unknown. To 419 feet traverse.		
Concealed through loose drift to 150 feet traverse from mouth of tunnel.		

With the provisional upper limit used in the above description of the tunnel section, the Keyser consists of 39 feet of impure lumpy limestone, below, and 54 feet of shale and sandstone, above.

Palmerton, Carbon County (Mauch Chunk quadrangle).—The section at this locality comprises three parts, one in the sand quarry of the Alliance Stone Company, ¾ mile northwest of the village, a second in an abandoned limestone quarry southwest of the sand quarry, and the third connecting the two quarries.

I. Section in Sand Quarry Alliance Stone Company

The Alliance Stone Company operates a large quarry for the production of sand, ¾ mile northwest of the center of the village. The following section, (measured by C. K. Swartz, assisted by C. D. Swartz), is seen here.

Devonian System Palmerton sandstone	Thickness, feet Beds	Total
Heavy-bedded gray disintegrating sandstone; quarry face reported 130 feet high	130	136
Coarse gray somewhat conglomeratic sandstone, forming floor of quarry, 2 to 4 inches chert at bottom, N. 49° E., 11° NW	6	6
Thickness Palmerton sandstone exposed		136
Bowmanstown chert		
Gray sandstone with white chert in middle	3	13
White fossiliferous chert, base of exposure by powder house. The cherts, which are profusely fossiliferous, bear *Atrypa reticularis, Amphigenia* sp., *Chonetes deflectus, Eodevonaria arcuata, Leptaena rhomboidalis, Leptocoelia acutiplicata, Meristella* sp., *Oriskania* n. sp., *Schuchertella arctostriata, Schuchertella pandora* (c), *Schuchertella* sp., *Spirifer duodenarius* (c), *Spirifer macrus* var., with plications in general fewer than typical (usually about 10 on side, some 8 on side, a few young, 6 on side) profuse, *Spirifer manni?* (r), *Stropheodonta* sp., *Loxonema* sp. aff. *delphicolum, Platyceras* sp. aff. *P. conicum, Tentaculitis scalariformis,* cyrtoceroid	10	10
Thickness Bowmanstown chert exposed		13
Concealed.		

II. Section Between Limestone and Sand Quarries

The section was measured by a traverse which begins at the eastern end of the limestone quarry and just west of a conveyor belt carrying sand from the sand quarry. The point of beginning is the top of a massive 6-foot ledge of limestone forming the base of the Keyser at this place. The traverse extends thence 497.5 feet to a point 1 foot below the top of the Bowmanstown chert in the sand quarry. Six feet of the Keyser limestone and 12 feet of the exposed Bowmanstown chert are thus included in the traverse.

The strike and dip observed in the limestone quarry are N. 69° E., 52° NW.; strike and dip observed on floor of sand quarry N. 49° E., 11° NW.; used in calculation N. 69° E., 21½° NW. Bearings are magnetic 1936.[10]

Devonian System

Bowmanstown chert

	Thickness, feet	
	Beds	Total
Gray sandstone and interbedded white chert. Traverse ends, 497.5 feet, at base of this unit	1	13
Concealed on traverse but exposed in quarry	12	12
Thickness Bowmanstown chert exposed in quarry		13
Concealed, embracing beds from lower part of Bowmanstown chert to upper part of Keyser limestone inclusive		235

Devonian or Silurian System

Keyser formation

Concealed on line of traverse. Shaly arenaceous limestone exposed west of conveyor belt bears *Chonetes jerseyensis* (a)	6	12
Massive arenaceous limestone. Traverse begins at its top	6	6
Thickness Keyser limestone described		12

III. Section in Limestone Quarry Southwest of Sand Quarry

This quarry is situated 500 feet southwest of the sand quarry of the Alliance Stone Company, being at the head of a small street shown upon the U. S. topographic map. Its elevation is about 600 feet. The quarry, now abandoned, extends for a considerable distance east and west. The section described by F. M. Swartz (1929, p. 25) is seen near the western end of the quarry.

[10] Traverse distances and directions are as follows: To road 0-100 feet N. 79° E., along and west of road 100-360 feet N. 16° W., along foot path 360-450 feet N. 70° W.; to base exposure on traverse 450-497.5 feet N. 58° W.; an upright pipe of gas tank at 489.3 feet. Elevation at end of traverse 90 feet above beginning. Measured by C. K. Swartz, assisted by C. D. Swartz.

Devonian or Silurian System

Keyser limestone	Thickness, feet Beds	Total
Concealed. Thin-bedded impure very arenaceous limestone, weathering buff, bearing *Chonetes jerseyensis*	6	12
Thick-bedded impure limestone, weathering buff. bearing, at top, Chaetetoid Bryozoa (a), *Chonetes jerseyensis, Stenochisma deckerensis*	6	6
Thickness Keyser limestone exposed		12

Silurian System

Bossardville limestone		
Thin-bedded somewhat shaly limestone, *Leperditia* sp.	6	77
Medium-bedded light greenish-gray somewhat shaly limestone, laminated toward top; Bryozoa 1 foot below top at east end of quarry (aa)	6	71
Dense blue-gray limestone, upper 4 feet thin-bedded, *Leperditia* sp. (aa), lower 3 feet medium-bedded, *Stromatopora?*	7	65
Thin-bedded fine-grained blue shaly limestone	6	58
Concealed	15	52
Dark-blue limestone with impure shaly layers, Bryozoa?	6	37
Concealed, 3 feet hard irregularly bedded dense blue limestone in middle	12	31
Irregularly laminated limestone, the bedding planes with wavy surfaces suggesting large ripple marks. The upper surface of this bed forms the foot wall in most of the quarry	10	19
Concealed	4	9
Medium-bedded blue limestone, lower part laminated, *Leperditia* sp. (a).		
Concealed	5	5
Thickness Bossardville limestone described		77

The quarry affords the best exposure of the Bossardville limestone along the Lehigh River. Unfortunately it is being slowly filled with waste material.

The entire section described consists of the following divisions:

	Feet
Palmerton sandstone	136
Bowmanstown chert (exposed)	13
Concealed, Bowmanstown chert to Keyser limestone inclusive	235
Keyser limestone	12
Bossardville limestone	77
Total thickness	473

The following fossils were found loose in the Bowmanstown chert in a small quarry west of the Alliance sand quarry, on the south slope of Stony Ridge: *Eodevonaria arcuata, Leptocoelia acutiplicata, Schuchertella* sp., *Spirifer duodenarius,* and *Spirifer macrus* var.

Little Gap, Carbon County (Mauch Chunk quadrangle).—Little Gap is situated 5 miles northeast of Palmerton, where Buckwha Creek transects Stony Ridge. A State highway passes through the gap on the

west side of the stream. The cut along the highway affords a section extending from the Oriskany sandstone to the base of the Devonian black shale. The exposures of the Bowmanstown chert and Palmerton sandstone are unusually fine. The beds are overturned to the north.

The upper part of the Palmerton sandstone is duplicated by a fault near the north end of the cut, the interval between the sandstone exposures being concealed by chert talus. A paint ore mine of the Prince Manufacturing Company was formerly worked north of the sandstone, the tracks entering it being still visible beneath the level of the highway.

The traverse begins at the southwest corner of a high retaining wall and extends 6 feet to the southeast corner of the wall, thence 382 feet southward along the highway to the end of the exposure.[11]

	Thickness, feet	
	Beds	Total
Devonian System		
Fissile black shale, near north end of cut. It is succeeded at the north by an impure argillaceous limestone, 90 feet above the sandstone, bearing *Schuchertella variabilis*, numerous species of *Chonetes* including *C. hemisphericus, C. lineatus* var., *C. mucronatus*, cf. *C. scitulus, C. setiger, Strophcodonta* sp. Lower beds concealed.		
Onondaga limestone		
Hydraulic limestone underlain by carbonate iron ore and clay are reported in former mine workings at this point on west side of gap.[12] They are now represented on the surface by 8 feet of thick-bedded, discolored, drab, calcareous shale underlain by a hematitic bed. Higher beds concealed.		
Palmerton sandstone		
Massive gray sandstone, many beds conglomeratic, with pebbles to ¼ to 1 inch diameter. Many beds tend to disintegrate upon exposure. Base at 221 feet traverse, elev. 11.5 feet.[13]		152
Total thickness Palmerton sandstone		152
Bowmanstown chert[13a]		
Interbedded white chert and gray conglomeratic sandstone. *Amphigenia elongata*, loose	12	111
White, somewhat sandy chert, *Spirifer macrus* var. (aa), in lower 4 feet. Base due at 265 feet traverse, elev. 16 feet	23	99
Chert like overlying, but weathering rusty brown; bearing, 2 feet below top, *Eodevonaria arcuata, Meristella* sp., *Schuchertella arctostriata, Spirifer macrus* var. (aa). Base at 337 feet traverse, elev. 18.5 feet	56	76
Concealed, hematitic bed in middle; age of unit uncertain.[14]	20	20
Total thickness Bowmanstown chert		111

[11] Traverse 0-6 feet N. 87° E., 6-221 feet S. 3° E., 221-end S. 4° E. Elevation of southeast corner of retaining wall is made zero. Culvert at 3+03.5 feet. Strike used N. 77¾° W. (average seven observations). Bearings magnetic 1939.

[12] Private communication from Mr. H. E. Sander, President Prince Manufacturing Co., to whom the authors are greatly indebted. The clay is reported 6-18 inches thick.

[13] Dip used, 44° SW. (average seventeen observations). Dips on four best bedding planes average 47¼° SW.

[13a] Dip used to end of traverse 48½ SW. (average eight observations).

[14] Traverse S. 4° W. from station 3+37 feet to 3+43 feet gives concealed interval 4.5 feet. Traverse from 3+43 feet N. 88° W. 23 feet, to base exposed Oriskany, elev. 37.5 feet, gives concealed interval 15.5 feet. Total concealed, 20 feet.

Oriskany sandstone	Thickness, feet Beds	Total
Massive gray conglomeratic sandstone, pebbles up to ½ to 1 inch diameter, bearing indistinct *Spirifer arenosus* and numerous fossil cavities. Traverse 15 feet (estimated) S. 4° W.	11	19
Greenish-yellow disintegrating sandstone, some pebbles up to ¼ inch diameter. Bearing *Meristella* sp., *Spirifer* sp., base at 388 feet traverse, elev. 20.5 feet	8	8
Thickness Oriskany sandstone measured		19

Concealed.

The section is notable for the marked thickening of the combined Bowmanstown-Palmerton interval toward northeast.

Kunkletown, Monroe County (Wind Gap quadrangle).—Two sections are exposed in the vicinity of Kunkletown, one 0.8 miles southwest, the other 0.9 miles southeast of the village.

I. Section 0.8 Mile Southwest of Kunkletown

This section is seen west of the road leading from Kunkletown to Smith's Gap school. A number of small quarries are opened in the Bossardville limestone at this point, west of the road. Extensive openings have been made by the Universal Atlas Cement Company in the New Scotland chert south of the limestone quarries. The section described begins at the base of the westernmost limestone quarry and extends thence, 90° to the strike, to the chert quarries. Bearings magnetic 1938. Distances paced.[15]

Devonian System		
Oriskany sandstone	Thickness, feet Beds	Total
Numerous massive gray sandstone boulders lie loose, south of the chert quarries. Loose boulders on the south slope of the pit bear *Meristella lata, Spirifer arenosus, Spirifer murchisoni.*		
New Scotland chert		
Dense white chert fragments, loose, in extensive workings of the cement company. Thickness estimated. The chert bears *Eatonia medialis, Leptaena rhomboidalis, Schuchertella woolworthana* (a), cf. *Stenochisma formosa, Uncinulus vellicatus* var., *Platyceras?* sp.		40–50
Coeymans? to Keyser formations, inclusive		
Concealed, Traverse 300 feet, 90 degrees to strike		180
Silurian System		
Bossardville limestone		
Platy blue limestone, exposed in quarry. Upper beds shaly. Dip 35° to 40° SE.		83

Concealed.

[15] This section was brought to the attention of F. M. Swartz by A. B. Cleaves.

II. Section 0.9 Mile Southeast of Kunkletown

The Universal Atlas Cement Company has opened extensive quarries in the New Scotland chert on the south slope of Chestnut Ridge, southeast of Kunkletown. The workings follow the strike of the cherts for a long distance northeastward, near the letter N of the words Chestnut Ridge, on the U. S. topographic map. The following section was measured by C. K. Swartz, assisted by C. D. Swartz.

Devonian System

Oriskany sandstone	Thickness, feet Beds	Total
Numerous fossiliferous boulders of massive somewhat ferruginous gray sandstone lie loose, north of the chert pits, or in the pits and along the railroad tracks leading to the pits. They bear *Edriocrinus* sp., *Camarotoechia (Plethorhyncha) speciosa, Rensselaeria ovoides* (a), *Spirifer arenosus* (c).		
New Scotland cherts		
Numerous fragments of dense white chert lie loose in the workings of the cement company bearing *Edriocrinus pocilliformis, Eatonia medialis* (c), *Rhipidomella oblata, Schuchertella woolworthana* (a), *Spirifer cyclopterus, Stropheodonta (Leptostrophia) beckii, Uncinulus abruptus, Phacops logani.* Thickness estimated.....		40–50
Concealed.		

Saylorsburg, Monroe County (Wind Gap quadrangle).—Two sections were measured in this vicinity, one near Mt. Eton Church, southwest of Saylorsburg, the other at Saylorsburg.

I. Section Near Mt. Eton Church

Mt. Eton Church is situated on the crest of Chestnut Ridge 1.3 miles southwest of Saylorsburg, at a fork of the road shown on the U. S. topographic map. The road shown going east is now abandoned but still exists as a lane near the church. The section described crosses this lane 800 feet east of the highway.

The traverse begins at the top of a small quarry in the Bossardville limestone. It extends westward along the strike of the beds, S. 67° W., 100 feet, and thence northward N. 23° W., perpendicular to the strike, intersecting the lane at 328 feet traverse. It ends at 440 feet traverse at the bottom, approximately, of the Oriskany sandstone exposed on the hillside. All the rocks of the section are overturned toward the north.[16]

[16] Dip used 40° S., being that in the railroad cut at Saylorsburg. Altitude of station 1+00 feet of traverse taken as zero. First 100 feet were paced, remainder measured by tape. Bearings magnetic 1937. Measured by C. K. Swartz, assisted by C. D. Swartz.

	Thickness, feet	
	Beds	Total
Devonian System		
Oriskany sandstone		
Massive coarse gray conglomeratic disintegrating sandstone, exposed in a sand pit east of line of traverse. The dip was not ascertainable but, seemingly, is low, as shown by the width of outcrop along the highway. A distinct rise marks the base of this unit. Both the sandstone in place and numerous boulders, found upon the strike along the lane, are very fossiliferous, containing *Spirifer arenosus* (a), *Spirifer murchisoni* (a), *Rensselaeria ovoides* (c).		
New Scotland to Keyser formations inclusive		
Concealed. Top at 940 feet traverse, elev. 20 feet. Numerous fragments of white chert are seen near the top of this interval along the State road where they bear characteristic New Scotland fossils, including *Eatonia medialis, Rhipidomella oblata,* and *Schuchertella woolworthana*	203	203
Concealed. Traverse S. 67° W., 100 feet along strike. Altitude top made zero	0	0
Total thickness New Scotland to Keyser inclusive		203
Silurian System		
Bossardville limestone		
Platy blue-gray limestone, exposed in a small quarry, N. 67° E., 60° S. Concealed.	5	5

Numerous large quarries in the Bossardville limestone, worked farther northeast along the strike, show thicknesses of the Bossardville up to 50 feet, with dips at many angles; its limits are not exposed.

This section is important as showing an interval of about 200 feet from the top of the Bossardville to the base of the Oriskany sandstone.

II. Section at Saylorsburg

The following section was measured by F. M. Swartz in the cut of the Saylorsburg branch of the Lehigh and New England Railroad, now abandoned, just east of the Lake House in Saylorsburg. It embraces parts of the Palmerton sandstone and Bowmanstown chert. The top of the former and base of the latter are not exposed. The beds are overturned to the north.

	Thickness, feet	
	Beds	Total
Devonian System		
Palmerton sandstone		
Concealed.		
Heavy-bedded gray finely conglomeratic sandstone; some large sandstone blocks included in top of sandstone as measured, may not be in place. Dip 46° S.	60	67
Medium-bedded fine-grained sandstone, some fine conglomerate, grains up to ⅛ inch diameter, weathering purplish. Seen on bank west of tracks	7	7
Total thickness Palmerton sandstone described		67

Bowmanstown chert	Thickness, feet Beds	Total
Medium-bedded impure chert, weathering white to buff and becoming porous. A few quartz pebbles up to ⅛ or rarely ½ inch diameter. Partially concealed. Twenty feet below top occur *Eodevonaria arcuata* loose. In lower 3 feet occur *Amphigenia* cf. *elongata, Eodevonaria arcuata, Leptocoelia acutiplicata, Meristella* sp., *Spirifer macrus* var. (aa), *Schuchertella* sp., *Proetus* sp.	33	77
Medium- to thin-bedded impure chert or siliceous, fine-grained shale, weathering to a pinkish, buff, or whitish color. More fractured than overlying beds. Lower part well exposed on bank east of tracks. Fossils rare	20	44
Siliceous shale to impure chert, like above, partially concealed	24	24
Thickness Bowmanstown chert described		77

Concealed to south.

This section is of interest since it shows the persistence of the Palmerton sandstone and Bowmanstown chert eastward in Cherry Ridge to within 3 miles of the thick Esopus shale of Godfrey Ridge to the north.

SECTIONS IN NORTHERN RIDGE

Bossardville, Monroe County (Wind Gap quadrangle).—Sections are exposed at two localities in this vicinity, in the village and on Godfrey Ridge 1½ miles northeast of the village. Both were measured by C. K. Swartz assisted by C. D. Swartz.

I. Section in Village

Bossardville is interesting as the type locality of the Bossardville. Several quarries have been opened here. The following section is exposed in the largest quarry, situated north of the highway, in the village.

Silurian System Bossardville limestone	Thickness, feet Beds	Total
Dark-blue laminated limestone, weathering thin-bedded at top, seen in north quarry wall. Dip 38° N. Vertical elevation 25 feet	20	86
Beds same as above, seen at west end of quarry	30	66
Dark-blue limestone, some green shale, in footwall of quarry	2	36
Limestone similar to overlying, but some beds seem less pure. Concealed in part beneath road. Dips 45° N. to 85° N.; 65° N. used. Traverse across beds, 37 feet 90 degrees to strike, descending 4 feet	34	34
Thickness Bossardville limestone measured		86

Both overlying and underlying beds are concealed.

II. Section on Road Across Godfrey Ridge

A road leading toward Stroudsburg crosses Godfrey Ridge 1½ miles north of Bossardville. Both the Oriskany sandstone and Esopus shale are seen on this road. The Oriskany is exposed at the beginning of the

steeper slope, in ascending the ridge, where it bears *Spirifer arenosus*. The Esopus shale is well exposed immediately above it to the north, bearing *Leptocoelia acutiplicata* (a) in its lower part and the same species associated with a few *Chonetes lineatus?* in higher beds. The thickness of the Esopus could not be measured at this point, but it is large and the beds are similar lithologically to those farther eastward.

The Esopus is also present, along the road crossing the western end of Godfrey Ridge to Kellersville, 2 miles farther southwest. There is no trace of the Palmerton sandstone in the sections along either of these roads, although it is fully developed at Saylorsburg, 5 miles away.

Stormville, Monroe County (Delaware Water Gap quadrangle).—The section at Stormville is of unusual interest, since it is from this locality that I. C. White derived a number of formational names long used in the stratigraphic geology of eastern Pennsylvania. These terms include, in ascending order, the Stormville hydraulic cement bed, Stormville limestone, Stormville conglomerate, and Stormville shale. The significance of these terms has long been open to question. It hence seemed desirable to re-examine White's type section to determine the meaning of names more exactly.

White unfortunately does not locate his sections very clearly. He does say, however, (White, 1882, p. 278, 281) that beds 3 to 9 of his section at Stormville were measured "opposite J. Snover's house." He speaks also (p. 1400) of the "stone house," which is evidently the same dwelling. An old stone house still stands at the east end of Stormville which old residents of the village state was the former residence of J. Snover. It is located at the intersection of the road to Delaware Water Gap and Pennsylvania highway 612, to Stroudsburg. A stream descends the hill at this point and crosses the highway a few feet west of the intersection.

The only section now available for measurement in the immediate vicinity is exposed along or in the bed of this stream. The section is much concealed. Its lower part crosses a minor anticline. It terminates above in a sharp syncline. The measurements here given were made in the north limb of this anticline.

The traverse begins on the east side of the stream, 180 feet N. 15° W. from the center of the valley road, and ascends the hill northwestward, along or near the stream, for a distance of 228 feet to the end of the section.[17]

[17] Measured by C. K. Swartz assisted by C. D. Swartz. Measurements to tops of beds. Intersection of roads taken as 455 feet above sea level from topographic map. Traverse 0 to 88 feet, N. 15° W.; 88 feet to 173 feet, northeast on strike; 173 feet to end, northwest, 90 degrees to strike.

Devonian System

Coeymans limestone

	Thickness, feet Beds	Total
Massive gray conglomeratic sandstone, forming crest of small ridge. N. 68° E., 60° NW., elev. top 547 feet. *Gypidula coeymanensis* (a), in some beds. The Stormville conglomerate. To 228 feet traverse	14	80
Massive gray arenaceous limestone, partially concealed. *Gypidula coeymanensis* (aa) in some beds	44	66
Concealed. Top at 173 feet traverse, elev. 525 feet.		
Concealed. Boulders massive calcareous sandstone, loose. Age questionable. Top at 88 feet traverse, elev. 526 feet	22	22
Approximate thickness Coeymans limestone		80

Devonian or Silurian System

Keyser Group

Manlius limestone

Concealed. A flat at top. Limits questionable. To 71 feet traverse, elev. 511 feet.	29	29
Approximate thickness Manlius limestone[18]		29

Rondout limestone

Fine-grained dove-blue dense argillaceous limestone. Some beds weather yellow. The foot of upper falls in stream is at 28 feet traverse. Top at 38 feet traverse, elev. 509 feet. The Stormville hydraulic cement rock	12	14.5
Laminated blue limestone, *Leperditia* sp. (a). Top at 25 feet traverse. White's Decker's Ferry limestone. This limestone is also exposed on the west side of highway route 612, about 100 feet above the intersection of the roads	2.5	2.5
Approximate thickness Rondout limestone		14.5

Decker sandstone

Blue-gray sandstone, upper foot largely concealed. Top at 22 feet traverse. This sandstone also outcrops as a massive conglomeratic ledge, 18 inches thick, on the highway, beneath the *Leperditia* limestone bed, referred to above. Its strike and dip at the latter place are N. 68° E., 30° SE.	6	27
Arenaceous limestone. Top is at the foot of a small falls at 17 feet traverse.	2	21
Concealed. Top at 15 feet traverse, elev. 503 feet	17	19
Argillaceous limestone, some pebbles present. N. 53° E., 59° N. Limestone bears *Camarotoechia litchfieldensis* (a), *Atrypa reticularis*. Elev. top 490 feet. The base of this bed is the zero of traverse	2	2
Thickness of Decker sandstone described		27

Concealed.

The Stormville conglomerate is overlain, according to White, by the Stormville shale to which he assigns a thickness of 160 feet. The latter

[18] The reference of these beds to the Manlius is based upon the similarity of this section to others in the northern ridge farther east.

is succeeded, in turn, by the Oriskany sandstone. The Stormville shale crops out in a belt nearly 1000 feet wide, measured perpendicular to the strike. It is largely concealed, and its strikes and dips are unknown. It is bounded by beds with high dips. It is manifest that the thickness of these beds cannot be as little as 160 feet unless there is much concealed minor folding rendering all measurements untrustworthy. Six miles northeast of Stormville this interval is occupied by four geologic horizons with a combined thickness of about 300 feet.

These facts permit the following correlations:

White's section	Authors' correlations
Oriskany sandstone, 25 feet exposed.	Oriskany sandstone, 25 feet exposed.
Stormville shale, 160 feet, concealed.	Concealed 300 feet, consisting elsewhere of Oriskany shale. Port Ewen shale. Becraft limestone. New Scotland limestone and shale.
Stormville conglomerate 45 to 47.5 feet, intertonguing with the Stormville limestone at the northeast.	Coeymans limestone, 80± feet (58 feet exposed) including 14 feet conglomeratic sandstone at top, (Stormville conglomerate).
Stormville limestone, 50 feet. Thickness combined Stormville conglomerate and Stormville limestone 95 to 97.5 feet.	Manlius limestone, 29± feet. Thickness combined Coeymans and Manlius limestones 109± feet.
Stormville hydraulic cement bed, 8 feet.	Rondout limestone, 14.5 feet consisting of Argillaceous limestone, 12 feet.
Decker's Ferry limestone, *Leperditia* sp. (a), 6 feet. Concealed 10 feet.	Laminated blue limestone, *Leperditia* sp. (a), 2.5 feet.
Decker's Ferry sandstone, 20 feet exposed.	Decker sandstone, 6 feet exposed, (19 feet on Hartman farm).
Concealed.	Concealed.

The correlation of the two sections is apparent.

Delaware Water Gap, Monroe County (Delaware Water Gap quadrangle).—Two fine sections are seen in this vicinity, one at Croasdale Manor, north of Water Gap village, the other in the cuts of the N. Y. S. & W. R. R. at Experiment Mills, ½ mile farther north. The older beds are exposed at the first locality, the higher at the latter place. Both sections are unequaled in Monroe County.

I. Section at Experiment Mills

This section is exposed in the cuts of the N. Y. S. & W. R. R. just west of Marshall Creek, opposite the mills of the Minisink Paper Company. The traverse begins at the west abutment of the railroad bridge over the highway leading to Shawnee and extends thence westward 1610 feet

along the north rail of the track. Bearings magnetic 1937, traverse changes determined by angles; measured by tape.[19]

The formations from the Coeymans to the Port Ewen were studied by F. M. Swartz (1939, p. 87-89), who has described them more fully elsewhere. The results here given are based upon that work. Only a brief summary is presented here.

The Oriskany and Esopus were studied by both authors.

	Thickness, feet	
	Beds	Total
Devonian System		
Onondaga limestone		
Concealed, loose limestone boulders north of the track. A limestone quarry situated west of the road leading to East Stroudsburg, its base essentially on the same strike, exposes the limestone more fully.		
Esopus shale[20]		
Thick-bedded bluish-gray arenaceous shale, schistose, weathering greenish, bearing *Leptocoelia acutiplicata* (c), *Spirifer* sp. aff. *tullius*, 248 feet above base (1433 feet traverse); *Leptocoelia acutiplicata* 50 feet below top. Top at 1597 traverse		275
Oriskany Group		
Oriskany sandstone[21]		
Interbedded black chert and fine-grained sandstone. Top at 833 feet traverse. Two feet below top occur: *Platyorthis planoconvexa, Rensselaeria* sp., *Schuchertella becraftensis, Spirifer arenosus, Spirifer murchisoni* (aa), *Stropheodonta?* sp.	6	60
Massive coarse-grained calcareous sandstone, somewhat conglomeratic, pebbles up to ¼ inch diameter, some thin lenses black chert	12	54
Heavy-bedded fine-grained sandstone, and black chert, chert predominates, some beds coarse-grained sandstone	42	42
Total thickness Oriskany sandstone		60
Oriskany shale		
Thick-bedded dark-gray to black siliceous shale, with lenses of dark chert in some beds, especially in lower 2 feet. Beds weather dark ash gray to brownish gray, tending to become iron-stained near base. Beds much fractured across bedding planes, more shaly parts exhibit a marked slaty cleavage. Average strike and dip N. 87° E., 79° W. (Strike average four observations, dip average nine observations.) *Leptocoelia flabellites* and *Spirifer* sp. aff. *murchisonia* abundant in upper 8 feet.[22] Top at 744 feet traverse		54
Total thickness Oriskany shale		54

[19] Traverses as follows, in feet: Coeymans 0-341 NW., New Scotland 341-395 NW., 395-465 N. 49° W., Becraft 465-485 N. 46° W.; Port Ewen 485-666 N. 46° W.; Oriskany shale 666-744 N. 46° W. (whistling post at 710 feet), Oriskany sandstone 744-833 N. 48° W.; Esopus grit 833-1233 N. 59° W.; 1233-1597 N. 65° W. (mile post JC 98 at 1404 feet. culvert at 1597 feet).

[20] Strikes and dips used in calculating thickness, 833 to 1233 feet traverse N. 87° E., 53½° W., 1233 to 1597 feet traverse N. 82.3° E., 36.2° NW. (average six observations).

[21] Average strike and dip used in calculating thickness N. 86¾° E., 75½° N. (Strike average 11, dip 12 observations.)

[22] *Spirifer* sp. aff. *murchisoni* and *Spirifer* n. sp. (a) are also found in still lower beds exposed along cement road leading from Delaware Water Gap village across Godfrey Ridge to Stroudsburg.

Helderberg Group

Port Ewen shale	Thickness, feet Beds	Total
Dark-gray somewhat siliceous shale, weathering ash gray to brownish gray. These beds are somewhat more resistant to weathering than are the underlying ones. Calcareous lenses, 1 to 1½ feet below the top, are highly fossiliferous bearing *Chonetes aroostookensis, Chonetes* n. sp. (aa), *Spirifer concinnus* (c), *Dalmanites dolphi, Dalmanites* sp., *Homalonotus vanuxemi* (a), *Thlipsurella multipunctata* (aa), Ostracoda of many species (aa). Top is at 666 feet traverse	29	136
Concealed. Some ash-gray shale with slaty cleavage exposed at 608 to 611 feet traverse. Top is at 630 feet traverse	30	107
Dark-gray somewhat calcareous shale, weathering ash gray, with oblique schistosity, breaking into rather soft, slaty pieces. Exposed on bank above tracks. Top projects to track at 592 feet traverse. Many fossils	38	77
Mostly concealed; some calcareous shale in lower and upper parts; oblique schistosity, weathering ash gray. Top at 539 feet traverse	39	39
Total thickness Port Ewen shale		136

Becraft(?) limestone		
Hard dark-gray impure limestone to siliceous calcareous shale, weathering ash gray. Breaking across bedding. Dip about 50° NW. Top at 485 feet traverse. Age insecure		14

New Scotland formation		
Calcareous shale. Mostly concealed. Shale weathers ash gray, splitting across bedding. Top at 465 feet traverse	43	76
Thick-bedded dark-gray siliceous, argillaceous limestone. Some layers of small nodules of dark chert, and rows of limestone nodules, up to 8 inches thick, weathering light gray and contrasting with the dark-brownish surface of the matrix. Dip 37½° NW., N. 53° E., 30½° NW. Concealed 11 to 17 feet above base, *Eospirifer macropleura*. These beds might be considered to represent the Kalkberg limestone of Chadwick. Top at 395 feet traverse	33	33
Total thickness New Scotland formation		76

Coeymans limestone		
Thick-bedded hard arenaceous gray limestone with lenticular beds of course sandstone and some small pebbles, especially near its base. The sand weathers in relief upon the surface. Concealed 6 to 9 feet above base. The Stormville conglomerate of I. C. White. Top projects to the track at 341 feet traverse	14	77
Thick-bedded hard somewhat arenaceous bluish-gray, finely to coarsely crystalline limestone, breaking into rather irregular fragments. Beds of dark chert 3 to 4 inches thick at 9 and 14 feet, some irregular nodules 13 and 17 feet, and a prominent 6-inch bed 24½ feet above base. Numerous 2- to 6-inch coquinas, composed largely of the valves of *Gypidula coeymanensis* in lower 8 feet and 10 to 13 feet above base. This species is somewhat less common at higher levels. The top of this unit is at 306 feet traverse, its base at 156 feet	28	63
Axis of anticline at 156 feet traverse. The lower beds are repeated east of axis.		
Concealed to base of Coeymans, about	35	35
Total thickness Coeymans limestone about		77

II. Section at Croasdale Manor

Croasdale Manor is situated 0.6 mile north of Cherry Run, on the road leading from Delaware Water Gap, via Experiment Mills, to East Stroudsburg. Two quarries in the Bossardville limestone were formerly worked here, one just west of the manor house, the other some 300 feet farther north. The exposures in the north quarry appear to be the actual type section of White's Bossardville limestone, although it was named from Bossardville village. Both quarries are overlain by the massive Decker sandstone. Measurements of Decker sandstone at the north quarry are by F. M. Swartz (1929, p. 24); other beds by C. K. Swartz assisted by C. D. Swartz.

1. Section at south quarry

The highest beds exposed here are seen just north of a small gully entering the north end of the south quarry N. 43° W. from the manor house (magnetic 1938). They overlie the upper Decker sandstone. Traverse 90 degrees to strike.

Devonian or Silurian System

Keyser Group

Rondout limestone	Thickness, feet: Beds	Total
Concealed in gently sloping field to the northwest. Fragments of yellow limestone of "pethstone" type are found at the bottom of this interval a few feet northeast of the section. White's Stormville cement bed.		
Laminated blue limestone, weathering gray. Dip 25° NW. *Leperditia* sp. abundant in this bed 45 feet northwest of traverse. White's Decker's Ferry limestone	3	11
Concealed for 20 feet	8	8
Thickness Rondout limestone measured		11

The section is continued west of the south quarry.

Decker sandstone

Massive gray sandstone, pebbles up to 1 inch diameter. Dip 18° W. Top 105 feet traverse, elevation 62 feet	48	90
Concealed. Top 67 feet traverse, elevation 23 feet	42	42
Total thickness Decker sandstone		90

Silurian System

Bossardville limestone

Thick-bedded impure limestone weathering yellowish, at top of quarry. Top of this bed is made zero traverse, zero elevation		10

Underlying beds, exposed in quarry, not measured.

2. Section at north quarry

Devonian or Silurian System

Keyser Group

Decker sandstone	Thickness, feet Beds	Total
Concealed.		
Massive conglomeratic gray sandstone, bearing *Spirifer modestus*....	36	82
Concealed.....	21	46
Massive conglomeratic gray sandstone, with many cavities elongated parallel to the bedding.....	11	25
Thin-bedded sandstone and arenaceous shale in quarry cliff; partially concealed.[23] A brown calcareous sandstone crops out, seemingly about this horizon, south of the quarry, where it bears *Chonetes jerseyensis* (aa).....	14	14
Total thickness Decker sandstone.....		82

Silurian System

Bossardville limestone

Thick-bedded impure limestone, weathering yellowish. Not accessible for examination. Questionably referred to the Bossardville..	10	91
Thin-bedded blue limestone above, calcareous shale at bottom.....	9.5	81
Massive dark-blue laminated limestone, *Leperditia* sp. Quarry rock	51.5	71.5
Concealed. Not improbably like overlying beds.....	1.5	20
Blue-gray shaly limestone, crinkly stratification, oblique schistosity, mud-cracked surfaces, producing prismatic columns, *Leperditia* sp. (a). N. 65° E., 33° SE. (average two observations).....	18.5	18.5
Concealed.		
Thickness Bossardville limestone measured.....		91

This section may be summarized as follows: upper impure limestones 19.5 feet, purer quarry rock 51.5, basal crinkly limestone and concealed 20 feet. White makes the upper beds 20 feet, quarry rock 65 feet, basal prismatic and concealed 25 feet.[24]

Shawnee-on-the-Delaware, Monroe County.—Shawnee-on-the-Delaware is situated 2 miles northeast of Delaware Water Gap village. Two fine sections are exposed in this vicinity, one south of the village, the other at Mosier Knob north of the village.

[23] I. C. White (1882.P.241) describes a greenish shale, 12 feet thick, in this position, stating that it is not fossiliferous.

[24] The Bossardville beds of this quarry were measured by running a traverse down the south face of the quarry with a steel tape, N. 29° E., to the bottom of the face, thence N. 14° E. upon the floor of the quarry to the prismatic beds which were measured directly, correcting for strike and dip.

I. Shawnee Village

The Rondout, Manlius, Coeymans and basal New Scotland formations are well exposed along the highway leading to Stroudsburg and the Delaware Water Gap, along the first hill south of the village. Magnesian limestone, apparently the "pethstone" of Cook, is seen about 360 feet south of the tradesman entrance road of the Buckwood Hotel; the higher beds are almost continuously exposed along the rise to the south. The section was measured by F. M. Swartz.

Devonian System

Helderberg Group

New Scotland shale and limestone	Thickness, feet Beds	Total
Concealed to south.		
Thick-bedded calcareous shale or shaly limestone, tending to fracture across bedding; with 6-inch lenticular beds of black chert at base, 2½ feet and 4½ feet above base. *Eospirifer macropleura* fairly common at 5 feet. Top reaches road level about 130 feet south of the culvert by the Stormville conglomerate	6.5	6.5
Coeymans limestone		
Somewhat pebbly gray limestone, less arenaceous than the next lower beds	4	79
Thick-bedded hard very arenaceous and, in part, pebbly limestone, grading into calcareous sandstone and fine quartz pebble conglomerate. The quartz pebbles range up to about ½ inch in greatest diameter. Exposed at and near a culvert 30 feet south of a private entryway on west side of road. Umbonal parts of ventral valves of *Gypidula coeymanensis* common to abundant in weathered residue in middle and basal part. These beds and the next overlying 4 feet of strata form the Stormville conglomerate of I. C. White	12	75
Thick-bedded hard bluish-gray, rather finely crystalline, irregularly arenaceous limestone, the sand grains generally small. The lower 2 feet of the Coeymans and the uppermost Manlius are exposed near a culvert; the lower 6 feet well exposed about 110 feet south of the culvert. These beds tend to split across the bedding and break up into irregular fragments. Where weathered, bands or lenses of purer limestone tend to show up against the mud-stained arenaceous crust of the remainder. There are beds or lenses of dark chert, as follows: a 2-inch thick lense at 13 feet; 2-inch nodules or lenses at 38 and 44.5 feet; top of a prominent 8-inch bed at 48 feet. *Gypidula coeymanensis* is rare in the lower 4 or 5 feet but becomes more abundant at higher levels, forming well-marked coquinas at 13 feet, and 6- to 12-inch lenticular coquinas from 19 to 28 feet. The upper 4 feet is seen in bank at private entryway to house on northwest side of highway. When much weathered these uppermost beds show their very arenaceous character, so that it is difficult to separate them from the overlying Stormville pebbly sandstone beds	63	63
Total thickness Coeymans limestone		79

Devonian or Silurian System Keyser Group Manlius limestone	Thickness, feet Beds	 Total
Thick-bedded hard dark-gray, in part very arenaceous, finely crystalline limestone. Weathered surfaces show edges or sections of valves of *Stropheodonta* sp., and some Stromatoporoidea. Upper 4 feet exposed in bank just above a highway culvert. Exposures are not very satisfactory for about 70 feet south of the culvert, but hard arenaceous limestone with numerous *Stropheodonta* sp., apparently the continuation of this bed, is exposed 70 to 120 feet south of the culvert[25]	5.5	31.5
Thick-bedded hard gray arenaceous limestone, with irregular hematitic laminae. Limestone much fractured, breaking into fairly large irregular fragments	2.5	26
Medium- to thick-bedded hard dark bluish gray finely crystalline arenaceous limestone	17	23.5
Concealed	6.5	6.5
Total thickness Manlius limestone		31.5
Rondout limestone		
Impure magnesian limestone, weathering light buff and punky, the weathered surfaces laminated parallel to the bedding (exposed 360 south of tradesman entrance to hotel), the "pethstone" of Cook	7.5	7.5
The underlying beds are concealed at this point.		

This locality affords the most complete section of the Coeymans and Manlius limestones observed in the area.

II. Mosier Knob

A fine partial section of the Oriskany sandstone is exposed on Mosier Knob, 0.4 mile north of the stream crossing at Shawnee village.

The beds, which are quite conglomeratic, contain, 40 feet below the top of the upper conglomerate, *Meristella lata?* (a), *M. rostellata* (a), *Rensselaeria* ? sp., and *Spirifer murchisoni.* Forty-seven feet below top of upper conglomerate occur *Leptocoelia flabellites, Spirifer murchisoni.*

Decker's Ferry, Monroe County (Wallpack quadrangle).—Decker's Ferry is of unusual interest since it is the type locality from which White (1882, p. 137-141) derived the names of three formations, the Decker's Ferry sandstone, limestone, and shale. Only the first of these names, however, is in use today.[26]

The Delaware River has cut a deep S-shaped curve through Wallpack Ridge at this point, forming bold, cliff-like exposures. Three sections were studied here, the most extensive on the road leading from the River

[25] The two culverts referred to are about 500 feet apart.

[26] White's Decker's Ferry limestone is of Rondout age, his shale of Bossardville (?) age, while his Decker's Ferry sandstone is distinct, lithologically and faunally, from both. The U. S. Geological Survey has introduced the term Decker limestone for the latter formation, a rather unfortunate name, since White used it for the quite different Rondout limestone.

Road to Bushkill, the second at Decker's Ferry, the third at the Wallpack Bend of the river. A fourth supplementary one was measured on Wallpack Ridge, one mile south of Decker's Ferry. (*See* Figure 2.) The sections will be considered in the order named.

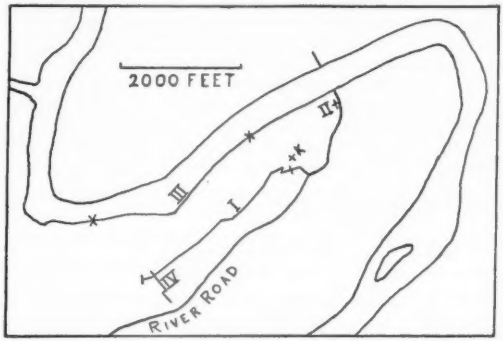

FIGURE 2.—*Map showing locations of sections in vicinity of Decker's Ferry*

I. Bushkill Road; II. Decker's Ferry; III. Wallpack Bend; IV. Wallpack Ridge; K. Limestone Knob.

I. SECTION ON ROAD LEADING TO BUSHKILL

A road leading to Bushkill leaves the River Road ½ mile south of Decker's Ferry and ascends Wallpack Ridge toward the southwest. Two studies were made along this road, one embracing the beds from the Rondout to the Coeymans, inclusive, the other from the New Scotland to the Oriskany sandstone. Measurements are in ascending order.

2. Section from New Scotland formation to Oriskany sandstone, inclusive

The traverse begins at the top of the exposed Coeymans on the road and ascends the ridge toward Bushkill. Bearings magnetic, 1936.[27]

[27] Descriptions of beds, traverses, and thickness of New Scotland, Becraft and lower Port Ewen are by F. M. Swartz. Elevations of stations and thickness of upper Port Ewen and Oriskany are by C. K. Swartz.

Traverse directions and distances in feet are as follows: New Scotland (including concealed Coeymans at base), 0-240 S. 65° W.; 240-470 S. 36° W., 470-683 S. 50° W.; Becraft 683-883 S. 52° W.; Port Ewen 883-918 S. 52° W., 918-1138 S. 65° W., 1138-1221 S. 76° W., 1221-1341 S. 79° W., 1341-1641 S. 54° W., 1641-1906 S. 57° W.; Oriskany shale 1906-2281 S. 67° W., 2281-2431 S. 60° W., 2431-2834 S. 61° W.; Oriskany sandstone and limestone 2834-2990 S. 88° W. Side traverse from 2760 feet.

Strike directions used in calculating thickness of upper shales are, stations 918-1341 N. 48° E., 1341-1906 N. 54½° E., 1906-2431 N. 59° E., 2431-2760 N. 61° E. Dip used, 918-2431 feet 35° NW.

Devonian System

Oriskany Group

Oriskany limestone and sandstone	Thickness, feet Beds	Total
Concealed under terrace deposits.		
Loose fragments siliceous limestone, exposed on hillside, containing *Spirifer murchisoni* (a), *Leptocoelia flabellites* (a) (variety with four or five plications on side). Top at 2990 feet traverse, elev. 658 feet		47
Oriskany shale		
Concealed, traverse along strike. Fragments of Oriskany siliceous or cherty limestone, loose, north of road. Side traverse from 2760 feet. Top at 2834 feet traverse, elev. 682 feet, at beginning of descent southwest.		
Concealed. El. 2281 feet traverse is 662 feet. Top at 2431 feet traverse, elev. 672 feet. Limits and age insecure		55
Helderberg Group		
Port Ewen shale		
Largely concealed. Some siliceous shale from base to 6 feet northwest of 1690 feet traverse. Top at 1906 feet traverse, elev. 644 feet.	15	174
Dark-gray to black, siliceous and some calcareous shale, cleaving across bedding; 6-inch bed impure limestone at 1341 feet traverse. Apparent strike and dip in lower part N. 42° E., 36° NW., (joint surface?). Top at 1641 feet traverse, elev. 639 feet	64	159
Concealed, top at 1221 feet traverse, elev. 602 feet	32	95
Dark-gray to black blocky siliceous to somewhat calcareous shale, cleaving across bedding; weathers into small, somewhat hackly, brownish fragments. *Dalmanites* sp. 10 feet above base (1055 feet traverse). Top at 1138 feet traverse, elev. 590 feet	32	63
Concealed. Top at 1028 feet traverse, elev. 573 feet	31	31
Thickness of beds referred to Port Ewen shale		174
Becraft limestone		
Medium to thick-bedded finely crystalline blue fossiliferous limestone, tending to break into small irregular fragments. Concealed 1 to 8½ feet above base; upper 5 feet may not be in place. N. 39° E., 31½° NW. Bearing, at base, *Atrypina imbricata* (r), *Eatonia singularis* (r), *Levenea subcarinata* (aa), *Rhipidomella* sp. (r), *Spirifer cyclopterus* (aa), *Spirifer perlamellosus* var. (a). Top at 883 feet traverse, elev. 563 feet		20
New Scotland formation		
Shale member		
Some calcareous shale near top, concealed below. Top at 683 feet traverse, elev. 563 feet. Side traverse from 654 feet traverse, elev. 555 feet	35	79
Calcareous to siliceous shale, much slaty cleavage. In lower 10 feet of traverse occur *Anoplia nucleata, Coelospira concava, Leptaena rhomboidalis* (a), *Meristella arcuata* (a), *M. laevis* (c), *Nucleospira ventricosa, Rhynchospira formosa?, Schuchertella woolworthana, Spirifer perlamellosus, Stropheodonta* (*Leptostrophia*) *beckii, Uncinulus vellicatus* var. Top at 470 feet traverse, elev. 539 feet	6	44

	Thickness, feet	
	Beds	Total
Much concealed, siliceous limestone and chert in part, bearing, at 320 feet traverse, *Eospirifer macropleura, Meristella arcuata* (a), *Spirifer cyclopterus, Stropheodonta (Leptostrophia) beckii, Trematospira multistriata;* at 255 feet traverse *Dalmanites pleuroptyx;* at 225 feet traverse *Bryozoa, Isorthis perelegans, Schuchertella woolworthana* (a), *Phacops logani.* Top at 338 feet traverse, elev. 517 feet	15	38
Limestone member		
Medium-bedded siliceous limestone, with some layers of chert, and lenses of light-gray limestone in darker matrix. The lower beds are exposed 62 feet N. 55° W. from zero of traverse; the upper 9-10 feet are seen on road between 147-172 feet traverse. Top at 172 feet traverse, elev. 505 feet	23	23
Total thickness New Scotland formation[28]		79

1. Sections from Rondout to Coeymans limestones, inclusive

Three sections were studied here, one south of the road, a second along the road, the third through a limestone knob north of the road. The beds in this vicinity are subject to such great variation in strike and dip, within short distances, that all the sections are necessary to secure trustworthy measurements of thicknesses.[29] The descriptions are based on work by C. K. Swartz, assisted by C. D. Swartz. Section (b) has also been described by F. M. Swartz (1939, p. 91).

a. Section southwest of Bushkill Road

The traverse begins at the top of the Decker sandstone which forms a cliff 350 feet S. 39° W. from station 0 + 00 of section on road. This point is best reached by going 183 feet S. 58° W. from station 1 + 50 on road, thence 154 feet S. 32° E. to the top of the Decker sandstone cliff. The measured section begins at the bottom of the cliff. This section gives, perhaps, the most trustworthy measurements of thicknesses of these beds.[30]

Devonian System
Helderberg Group
Coeymans limestone

	Thickness, feet	
	Beds	Total
Massive gray conglomerate sandstone, *Gypidula coeymanesis* (aa) in lower part. General strike N. 50° E. at northeast, N. 39° E. at southwest; used N. 45° E., 35° NW. The *Stormville conglomerate.* Top at 239 feet traverse, elev. 496 feet[31]	6	80.5

[28] The top of the Stormville conglomerate is exposed 140 feet S. 62° E. from station 6+54 feet on road, at elevation 550 feet; its strike and dip N. 46° E., 28½° NW., (average two observations), giving total thickness New Scotland 75 feet. The top of Stormville conglomerate is also exposed 335 feet S. 30° E. from station 27+60 feet on road, at elevation 539 feet, dip 48° NW. The general strike between stations 6+54 and 27+60 feet is N. 55° E. The strike angle increases from NE. to SW.

[29] Elevations are above sea level, the intersection of the River and Bushkill roads being taken as 410 feet from topographic map. Bearings magnetic, 1936.

[30] Traverse directions: 0-154 feet N. 32° W., 154-239 feet N. 6° W.

[31] A traverse was run from the top of the Stormville conglomerate connecting with section on the road as follows: 58 feet N. 30° W. to road, thence 207 feet N. 65° E., thence 30 feet S. 32° E. to station 2+88 feet at top of crinoidal limestone on road.

	Thickness, feet Beds	Total
Concealed	21.5	74.5
Concealed. Top at 154 feet traverse. Top is also 183 feet S. 58° W. from station 1 + 50 feet on road	39	53
Massive gray crinoidal limestone bearing, 3½ feet above base, *Favosites helderbergiae, Camarotoechia semiplicata, Gypidula coeymanensis* (aa); 5½ feet *Gypidula coeymanensis, Uncinulus mutabilis, U. pyramidatus;* at top *Favosites helderbergiae, Gypidula coeymanensis* (a). N. 55° E., 43° NW. (average three observations). Elev. top 492 feet	11	14
Heavy-bedded gray calcareous sandstone bearing in lower foot *Favosites helderbergiae, Gypidula coeymanensis, Strophonella punctulifera* (a); in upper 2 feet *Favosites helderbergiae* (a)	3	3
Total thickness Coeymans limestone		80.5
Devonian or Silurian System		
Keyser Group		
Manlius and Rondout limestones		
Dark-gray to black uneven-bedded argillaceous limestone, upper foot somewhat crystalline, bearing in upper 2 feet *Stropheodonta varistriata* (a). Top of Manlius limestone at 76 feet traverse	3	57.5
Concealed	54	54
Total thickness of Manlius and Rondout limestones		57.5
Decker sandstone		
Massive gray cross-bedded sandstone, flat solution cavities, conglomeratic 10-16 feet above base, N. 58° E., 44° NW., bearing *Chonetes jerseyensis.* Top made zero of traverse, elev. top 485 feet	19.5	22.5
Arenaceous limestone much concealed, bearing *Chonetes jerseyensis.*	3	3
Thickness Decker sandstone described		22.5

b. Section along Bushkill road

The traverse begins at the top of the laminated blue Rondout limestone, the zero point being 302 feet, measured along the Bushkill road, from the center of the River Road. The traverse extends thence westward 307 feet to the sharp curve of the road toward the south.[32]

Devonian System Helderberg Group Coeymans limestone	Thickness, feet Beds	Total
Concealed to approximate top of Coeymans limestone, possibly 6 to 8 feet	7	80
Dark fine-grained limestone, some black chert nodules, bearing *Gypidula coeymanensis.* Dip 27° NW., top at 307 feet traverse, elev. 485 feet	9	73

[32] Traverse directions: 0-175 feet S. 80° W., 175-288 feet N. 20° E., 288-307 feet N. 32° W. Strikes and dips used in calculations 0-175 feet N. 56° E. 39° NW., 175-288 feet N. 50° E. 36° NW. Estimates of thickness approximate.

	Thickness, feet Beds	Total
Massive gray crinoidal limestone, bearing at base *Gypidula coeymanensis* (a), *Uncinulus mutabilis;* middle of bed; *Gypidula coeymanensis* (a), *Uncinulus mutabilis;* 2 feet below top *Favosites helderbergiae, Gypidula coeymanensis* (a), *Uncinulus mutabilis;* 1 foot below top Cyathophylloid corals, *Atrypa reticularis* (a), *Gypidula* cf. *angularis,* G. *coeymanensis* (a), *Spirifer cyclopterus, Sp. octocostatus?, Stropheodonta* (*Leptostrophia*) *beckii.* Side traverse from station 2 + 82 feet traverse. Top at 288 feet traverse, elev. 484 feet	30	64
Concealed on road but exposed in cliff north of road. Top at 222 feet traverse, elev. 473 feet	16	34
One foot calcareous sandstone at base, concealed above along road. Top at 175 feet traverse, elev. 470 feet. Beds are exposed in cliff north of road	18	18
Total thickness Coeymans limestone approximately		80
Devonian or Silurian System		
Keyser Group		
Manlius limestone		
Irregularly bedded gray limestone, partially concealed, bearing, 1½ feet above base, *Stropheodonta varistriata* loose, *Megambonia aviculoidea* (a), *Leperditia* sp. (aa); 8 feet above base (76 feet traverse) *Spirifer vanuxemi;* at top *Stropheodonta varistriata* (a). Top at 114 feet traverse, elev. 467 feet	23.5	34
Concealed on road. Beds like the overlying are reported in this interval by F. M. Swartz (1939, p. 91), bearing *Tentaculites gyracanthus* (c) 6 to 10 feet above base. Top at 57 feet traverse, elev. 456 feet	10.5	10.5
Total thickness Manlius limestone		34
Rondout limestone		
Concealed	6	24.5
Argillaceous limestone, weathering to irregular, yellow fragments, White's Stormville hydraulic cement rock ("pethstone")	7	18.5
Blue limestone, upper part laminated, *Leperditia* sp. (aa) near and at top. White's Decker's Ferry limestone.		
The top of this bed is zero of traverse, elev. 445 feet	5	11.5
Concealed 15 feet horizontally	6.5	6.5
Total thickness Rondout limestone[33]		24.5
Decker sandstone		
Conglomeratic gray sandstone		1.5
Concealed.		

c. Section through knob north of road

A sharp knob is situated 200 feet northeast of the road. Its summit is 220 feet N. 60° E. from station 2 + 68 feet of preceding section. A cliff forms its east side.

[33] F. M. Swartz (1939, p. 91) measuring independently reports the blue limestone 8 feet thick, concealed (below) 5 feet, thickness combined Manlius and Rondout 59 feet.

The point of beginning is on the face of the cliff, 50 feet N. 31° W. from zero of section b, its elevation 468 feet. A section of part of the Coeymans and Manlius was made here. The Manlius-Coeymans contact was then followed 116 feet northeast, along the face of the cliff (giving a strike and dip of N. 60° E., 36° NW. by 3-point method). The section was continued through the crest of the knob, perpendicular to the strike, to the top of the west face of the knob, thence 60 feet N. 56° W. to the top of the Coeymans.

Devonian System
Helderberg Group
Coeymans limestone
Section west of knob

	Thickness, feet	
	Beds	Total
Boulders of siliceous limestone and conglomeratic sandstone loose, at sharp change of slope descending toward river. Top at 60 feet traverse, elev. 501 feet. Concealed to west	1	83
Concealed, black cherts in soil in upper part. Top at 59 feet traverse, elev. 502 feet	15	82
Ledges and boulders massive, crinoidal limestone. Top at 16 feet traverse, elev. 509 feet	—2	67
Section through crest of knob		
Massive gray crinoidal limestone forming knob, elev. 520 feet. Horizontal traverse (estimated) 40 to 50 feet. Used 40 feet	66	69
Massive gray calcareous sandstone, elev. base 462.5 feet	3	3
Total thickness Coeymans limestone[34]		83
Section at point of beginning		
Coeymans limestone		
Massive gray crinoidal limestone, forming top of ridge, thickness exposed approximately	20	26.5
Massive arenaceous limestone, upper 2 feet bearing *Favosites helderbergiae* (a)	3.5	6.5
Dark-gray calcareous sandstone, bearing, 6 inches above base, *Strophonella punctulifera*	3	3
Thickness Coeymans limestone exposed about		26.5
Devonian or Silurian System		
Keyser Group		
Manlius limestone		
Massive dark-gray limestone, irregularly bedded, bearing, 1½ feet below top, *Strophеodonta varistriata*. A marked change of slope occurs at top, elev. 475 feet		7
Concealed.		

The thickness from zero of section b, on road, to base of Coeymans in cliff, using strike and dip N. 60° E., 36° NW., is 54 feet, making interval from top of Decker sandstone to base of Coeymans 65 feet; to top of Coeymans 148 feet.

[34] General strike and dip at top of Coeymans is N. 50° E. 27° NW., in knob N. 60° E. 36° NW. Used in calculations N. 55° E. 31½° NW.

II. Section at Decker's Ferry

The section described is exposed on the bluff west of the former residence of P. H. F. Smith, a short distance south of the ferry. The measurements begin at the level of the road and ascend the hill to the west. The Rondout limestone is concealed.[35]

Devonian or Silurian System
Keyser Group
Decker sandstone

	Thickness, feet Beds	Total
Calcareous gray conglomeratic sandstone, quartz pebbles up to ¼ inch diameter	8	60
Arenaceous limestone, calcareous sandstone 10-20 feet above its base, beds much concealed. 10 feet above base *Atrypa reticularis* (a), *Camarotoechia litchfieldensis, Chonetes jerseyensis;* 12.5 feet above base *Chonetes jerseyensis* (a), 19 feet above base Bryozoa (a) *Atrypa reticularis, Chonetes jerseyensis, Stenochisma deckerensis*	35	52
Calcareous sandstone making a bold ledge on crest of ridge, very fossiliferous, 1 foot above base Bryozoa, *Cornulites* sp., *Atrypa reticularis, Camarotoechia litchfieldensis* (a) *Chonetes jerseyensis* (aa), *Stenochisma formosa, Uncinulus keyserensis, Loxonema?* sp.	6	17
Concealed	8	11
Massive dark-gray irregularly bedded limestone, bearing *Chonetes jerseyensis, Dalmanella* cf. *postelegantula, Dinorthis flabellites, Rhynchospira formosa, Stropheodonta* (*Leptostrophia*) *bipartita*	3	3
Total thickness Decker sandstone		60

Silurian System
Bossardville limestone

Impure straticulate thin-bedded blue limestone, weathering yellowish	1.5	22.5
Concealed (White's Decker's Ferry shale?)	14	21
Laminated dark-blue limestone, weathering gray, calcite streaks in lower part, N. 45° E., 41° NW., *Leperditia* sp. (a) in upper part	7	7
Concealed.		
Thickness Bossardville limestone described		22.5

III. Section at Wallpack Bend of Delaware River[36]

Devonian System
Esopus shale

	Thickness, feet Beds	Total
Thick-bedded siliceous shale, slaty cleavage. Large thickness exposed toward west end Wallpack bend, not measured.		
Concealed		20

[35] Measured by C. K. Swartz, assisted by C. D. Swartz. The independent measurement by F. M. Swartz (1939, p. 91) is very close to that recorded here.

[36] Measured by F. M. Swartz. The traverse begins at base of exposed Oriskany and descends in section as follows (distances in feet): Oriskany and Port Ewen shale 0-75 N. 88° E., 75-325 N. 86° E. 325-725 S. 88° E., 725-1000 S. 81° E., 1000-1025 N. 78° E., 1025-1300 N. 78° E., 1300-1750 N. 69° E., 1750-1925 N. 69° E., 1925-1975 N. 59° E.; Becraft 1975-2150 N. 58° E.; New Scotland 2150-2225 N. 64° E., 2225-2715 N. 49° E., 2615-2635 N. 40° E., 2635-3060 N. 58° E. Bearings magnetic 1938.

Oriskany Group

Limestone and sandstone member

	Thickness, feet	
	Beds	Total
Thick-bedded conglomeratic sandstone, probably somewhat calcareous, numerous quartz pebbles ⅛-¼ inch, rarely ½ inch diameter in upper part, rare or absent in lower part. N. 56° E., 39° NW.	4	68
Medium- to thick-bedded, irregularly bedded, black chert and some siliceous or finely arenaceous limestone; 6-inch beds of coarse sandstone or fine conglomerate 2 and 6½ feet below top	17	64
Thin- to medium-bedded blue-gray siliceous limestone, grading into impure, dark chert. *Spirifer murchisoni* (c). N. 51° E., 33° NW. Traverse begins at base of this unit	47	47
Total thickness Oriskany limestone and sandstone		68

Oriskany shale and Port Ewen shale

Concealed. Base of this interval at 1025 feet traverse. Calculated on basis of strikes and dips observed along river, this interval would measure about 325 feet.

Impure siliceous limestone to calcareous siliceous shale, concealed in large part. Dip at 1450 feet traverse, 22° NW. *Dalmanites* sp. loose at 1300 feet traverse. Base at 1750 feet traverse. Calculated on basis of strikes and dips observed along river, these beds would measure about 115 feet in thickness, and the total interval from the top of the Becraft limestone to the base of the Oriskany limestone and sandstone would measure 440 feet. This calculation is apparently much too great in view of the other measurements obtained in the vicinity and is not reliable in view of the small angle to the strike and of the possibility of minor flexures in the long concealed interval.

Becraft limestone

Thick-bedded dark-gray finely to, in part, rather coarsely crystalline limestone, seen mostly as loose boulders, but with some small exposures. There are several layers of 1- to 2-inch nodules of dark chert, and a few greenish, argillaceous laminae. The limestone tends to break into irregular, somewhat lumpy fragments. N. 42° E., 28° NW. Base about at 2150 feet traverse	23	23

New Scotland formation

Shale member

Concealed; base at 2225 feet traverse	6	55
Thick-bedded calcareous shale and impure limestone, exposed at and near rocks locally known as the "five loaves of bread." N. 49° E., 36° NW. Very fossiliferous, bearing *Rhipidomella oblata* (c), *Leptaena rhomboidalis* (c), *Strophеodonta* (*Leptostrophia*) *woolworthana* (a), *Eatonia medialis* (c), *Eospirifer macropleura* (a), *Spirifer perlamellosus* (c), *S. cyclopterus* (c), *Coelospira concava* (c), *Trematospira multistriata* (r), and other fossils. Base at 2615 feet traverse	49	49
Approximate thickness of shale member		55

Limestone member

Medium-bedded somewhat siliceous limestone with some layers of chert. *Isorthis perelegans* (r-c), *Schuchertella woolworthana* (r), *Eospirifer macropleura* (c), *Phacops logani* (r). These beds are well exposed, but their thickness was not measured. Below them is seen part of the Coeymans limestone, with 6 to 8 feet of calcareous, somewhat conglomeratic sandstone (Stormville conglomerate) at top.

IV. Section on Wallpack Ridge 1 Mile Southwest of Decker's Ferry

Because of the difficulty of securing trustworthy measurements for the Port Ewen and Oriskany shales in the preceding sections, a confirmatory section was measured here perpendicular to the strike. The traverse begins 100 feet northwest of the 2760 feet station of the Bushkill road traverse, then extends southeastward, through station 2760 feet, to the top of the Bossardville limestone.[37]

	Thickness, feet	
	Beds	Total
Devonian System		
Oriskany Group		
Limestone and sandstone member		
Concealed under terrace deposits.		
Concealed. Loose fragments siliceous limestone on slope contain *Spirifer murchisoni* (a), *Leptocoelia flabellites* var. (a). Top at zero traverse, elev. 658 feet, base at center of road, 100 feet this traverse. Elevation center of road 682 feet.[38]	47	47
Oriskany shale to New Scotland formation, inclusive		
Concealed above. At base, dark shaly limestone and calcareous shale, with lighter areas, New Scotland limestone member, 15 feet thick. Base at 435 feet traverse, elev. 539 feet	325	325
Coeymans limestone		
Massive calcareous sandstone, forming a prominent ledge on hillside. Top at 435 feet traverse, elev. 539 feet, dip 48° W. The Stormville conglomerate, consisting of cross-bedded conglomeratic calcareous sandstone, 2½ feet thick, at top, underlain by crystalline limestone with some black chert, 2½ feet, and massive gray calcareous sandstone, with *Gypidula coeymanensis* (a) in its upper part, 3 feet	8	8

The section is continued from the top of the ledge formed by the Stormville conglomerate, 100 to 150 feet farther southwest, and then descends the hill perpendicular to the strike. It reaches the River Road 0.6 mile south of the intersection of the River and Bushkill roads, opposite a barn on the property of the Penn. Power and Light Co. (formerly J. Van Auken).

	Thickness, feet	
	Beds	Total
Coeymans limestone		
Massive gray calcareous conglomeratic sandstone; 10 inches chert in middle. *Gypidula coeymanensis* (a) in lower part. Dip 55° NW. The Stormville conglomerate	14.5	41.5
Slope surface 39° SE.		
Massive gray crinoidal limestone; *Gypidula coeymanensis* (a) in some beds. Top 133 feet, slope measurement, 84 feet vertically above top Decker sandstone	27	27
Thickness Coeymans described		41.5

[37] Measured by C. K. Swartz, assisted by C. D. Swartz. Dip used in calculation, 40° NW.

[38] Ledges of fossiliferous, cherty limestone, 6 feet thick, are exposed on the crest of the ridge 800 feet farther southwest, the northern slope of the ridge being concealed. A careful section made at this point embracing the beds from the top of the Stormville to the top of the exposed Oriskany gave a thickness of 372 feet, the same as here.

Coeymans, Manlius, and Rondout limestones	Thickness, feet Beds	Total
Concealed. Top 106 feet, slope measurement, 67 feet vertically above top Decker sandstone	106	106
Decker sandstone		
Massive gray sandstone, exposed a short distance north of line of traverse. Dips up to 52° NW., variable (creep?)	12	63
Concealed in large part, a short distance north of line of traverse, at base, 1½ feet limestone weathering yellow, carrying *Atrypa reticularis* (a), *Chonetes jerseyensis* (a)	31	51
Concealed, age uncertain, to approximate base Decker sandstone	20	20
Approximate thickness Decker sandstone		63
Silurian System		
Bossardville limestone		
Concealed	14	14
Heavy-bedded dark-blue limestone, quarry rock, dip 58° NW. (average three observations), its top 65 feet stratigraphically (60 feet horizontally, 29 feet vertically) below base of 12-foot bed of Decker sandstone. Limestone extending to road. Thickness not measured.		

The Coeymans, Manlius, Rondout interval, 147.5 feet, may be compared with 138 feet south and 148 feet north of the Bushkill road. The interval from the top of the Decker sandstone to the top of the Bossardville quarry rock, 77 feet, compared with 75 feet at Decker's Ferry shows that this sandstone is the upper Decker sandstone and is overlain by the Rondout limestone. The general agreement suggests reasonable accuracy in very difficult measurements.

OTHER SECTIONS

In addition to the sections here described, graphic sections are also shown at Fox Gap (Swartz and Swartz, 1931, p. 646) and at Poxono Island (White, 1882, p. 223).

The formations described can be traced through the area by their lithology and faunas, to Decker's Ferry in Pennsylvania, where they enter New Jersey. They were recognized and described, with their characteristic lithology and faunas, at Nearpass quarry, in northern New Jersey, by Weller (1903, p. 63-67) and at Trilobite Mountain, in southeastern New York, by Shimer (1905) as shown graphically on Plate 1. (*See also* Hartnagel, 1905).

DESCRIPTION AND CORRELATION OF FORMATIONS

EARLY MIDDLE DEVONIAN FORMATIONS AND THEIR RELATIONS TO THE ORISKANY GROUP AND ESOPUS SHALE

General statement.—The Oriskany sandstone, so well developed in central Pennsylvania, is absent, as already stated, over the western part

of the region between the Susquehanna and Schuylkill rivers, south of the coal measures, being cut out by an unconformity or, in part, by faulting. As one goes eastward from the Susquehanna River, south of the coal measures, a sandstone appears beneath the Onondaga formation near the Schuylkill River and increases in thickness eastward. At the Lehigh River the sandstone is tripartite, consisting of an upper division 125-160 feet thick, a lower conglomeratic sandstone 25 feet thick, and an intervening chert of variable thickness. The sandstone forms two ridges west of the Delaware River, a southern and a northern.

Description of formations in southern ridge.—The southern ridge extends under various names from the Little Schuylkill River to the vicinity of Saylorsburg. The sandstone is well developed in it.

PALMERTON SANDSTONE: The upper division of the sandstone has the lithologic appearance and, seemingly, the stratigraphic position of the Oriskany sandstone of central Pennsylvania. It has, in part, a calcareous cement, is iron-stained and friable at many places, and is locally conglomeratic.

Oriskany fossils have been reported from it by various workers, but the writers have never seen an identifiable species in it. None have been reported in place, so far as we know. The only organisms seen by us are a few casts of cyathophylloid corals, too poorly preserved to permit identification. Like the Oriskany of central Pennsylvania this sandstone is overlain by Onondaga beds, from which it has generally been thought to be separated by an unconformity.

This sandstone has been called the Oriskany sandstone by all previous investigators.[39] In order to speak of it more objectively it is here called the Palmerton sandstone, from fine exposures on the ridges and in the quarries north of Palmerton, Pennsylvania.

The sandstone in question is underlain by thick, shaly beds, suggesting the relations observed in the Oriskany of central Pennsylvania. Where best measured they are about 275 feet thick. They have been termed the Oriskany shales by Chance (1882, p. 356-357) but comprise deposits of different ages.

BOWMANSTOWN CHERT: The Bowmanstown chert immediately underlies the Palmerton sandstone, forming the upper part of the so-called "Oriskany shale." It consists of an impure, somewhat sandy chert, some white and some black. Interbedded with it are thin sandstones and arenaceous shales. This chert is found beneath the Palmerton sandstone

[39] The results here stated were communicated to A. B. Cleaves and Bradford Willard in 1937. In a recently issued work these investigators (1939, p. 153) generously state their indebtedness to the writers.

wherever its horizon is exposed in the southern ridge, from Bowmanstown eastward. It receives its name from exposures on the State highway one half mile east of Bowmanstown, where it is about 67 feet thick. It is over 100 feet thick at Little Gap.

The upper part of the Bowmanstown chert is profusely fossiliferous and contains beautifully preserved fossils. These include the following diagnostic forms: *Spirifer macrus* var. (aa), *Spirifer duodenarius, Amphigenia* cf. *elongata, Eodevonaria arcuata, Leptocoelia acutiplicata,* and *Schuchertella pandora.*

ORISKANY SANDSTONE: Beneath the Bowmanstown chert and forming a part of the "Oriskany shale" of former authors is a massive sandstone bearing characteristic Oriskany fossils. It is heavy-bedded, gray, and, at most localities, strongly conglomeratic. It bears fossils, locally in profusion, including *spirifer arenosus* (a), *Spirifer murchisoni* (a), *Camarotoechia* (*Plethorhyncha*) *speciosa, Meristella rostellata,* and *Rensselaeria ovoides.*

The Oriskany sandstone is of wide extent, being found at Schuylkill Haven and throughout the southern ridge, from the Lehigh River eastward. It is 25 feet thick at the Lehigh River.[40]

Correlations of formations in southern ridge.—The correlations of these formations will now be considered.

ORISKANY SANDSTONE: The basal formation in the above described sequence is the Oriskany sandstone. Its diagnostic species, including *Spirifer arenosus, Spirifer murchisoni, Camarotoechia* (*Plethorhyncha*) *speciosa, Meristella rostellata,* and *Rensselaeria ovoides,* are all Oriskany forms, found also in the Ridgeley sandstone, of Oriskany age, in central Pennsylvania and Maryland. The lithology of the beds is also similar to that of the Ridgeley sandstone with which these beds are clearly to be correlated, although they are more conglomeratic. *See also* Cleaves (1939, p. 97).

BOWMANSTOWN CHERT: The Bowmanstown chert, which immediately overlies the Oriskany sandstone, bears a considerable and well-preserved fauna. Its characteristic species, including *Spirifer macrus, Spirifer duodenarius, Amphigenia* cf. *elongata, Eodevonaria arcuata,* and *Leptocoelia acutiplicata,* are Middle Devonian forms. Many are also found in the overlying Onondaga limestone. None of the critical species have been observed in the Oriskany. The Bowmanstown chert is evidently of early Middle Devonian age.

[40] This sandstone was evidently united with the Palmerton and Bowmanstown by I. C. White (1882, p. 123) who made his Oriskany sandstone 200 feet thick at Bowmanstown. H. M. Chance (1882, p. 357) included it in the Oriskany shale.

PALMERTON SANDSTONE: No determinative fauna has been observed in the Palmerton sandstone. It overlies the Bowmanstown chert and underlies the Onondaga limestone; its age is, hence, between that of these formations. It is thus clearly post-Oriskany and not the Oriskany sandstone, as it has been made by preceding workers. It is early Middle Devonian.

NEW SCOTLAND CHERT: The Oriskany sandstone is underlain in the southern ridge by white chert bearing a profusion of New Scotland fossils.

SUMMARY OF SEQUENCE: The following sequence is thus seen in the southern ridge. (The thicknesses given are at the Lehigh River):

Middle Devonian

Onondaga limestone. Middle Devonian fauna. Thickness variable, increasing eastward.
Palmerton sandstone. Nonfossiliferous, 130 to 150 feet thick.
Bowmanstown chert. *Spirifer macrus* var. fauna, 60 to over 100 feet thick.

Lower Devonian

Oriskany sandstone. *Spirifer arenosus* fauna, 25 feet thick.
New Scotland chert. *Eospirifer macropleura* fauna, 40 feet thick.

The combined thickness of the Palmerton and Bowmanstown formations is about 200 to 250 feet.

Description of formations in northern ridge.—The Palmerton sandstone forms a ridge, shown in Figure 1, which extends, under various names, from a few miles west of the Lehigh River to within about 8 miles of the Delaware River where it vanishes. At the same time another ridge, involving beds of similar age, and here called the northern ridge, rises about 1½ miles northwest of the first ridge and parallel to it.

A surprising change occurs in passing from the southern to the northern ridge. The sandstone appears to be reduced in thickness from about 150 feet of almost nonfossiliferous beds to 55 feet of cherty, sparingly conglomeratic, richly fossiliferous sandstone, bearing many Oriskany species. At the same time a number of new or thicker units appear beneath it, while between the Oriskany and Onondaga appears the Esopus shale, nearly 300 feet thick. The ridge formed by these beds extends northeastward into southeastern New York, crossing the Delaware River at Decker's Ferry. I. C. White (1882, p. 130-131) was well acquainted with these changes and suggested that the Helderberg shales of the northeast was transformed into the Oriskany sandstone at the southwest, both being contemporaneous. The key to this situation lies in the fact that the Palmerton sandstone is not the Oriskany, with which it was confused, but a higher sandstone of Middle Devonian age.

ORISKANY SHALE: A new unit appears beneath the Oriskany sandstone of the northern ridge. It is an arenaceous shale, prevailingly dark, some-

what cherty. Its base has a tendency to become iron-stained upon weathering. It contains few fossils save in its upper 8 to 10 feet, where *Spirifer* sp. aff. *murchisoni* and *Leptocoelia flabellites* are common to abundant. *Spirifer* sp. aff. *murchisoni* and *Spirifer* n. sp. (a) are also found in the lower beds.

The Oriskany shale is best exposed in the cut of the N. Y. S. & W. R. R., at Experiment Mills, opposite the mills of the Minisink Paper Company, where it is 54 feet thick. Like the Oriskany sandstone it thickens and becomes more calcareous northeastward until it joins the Oriskany sandstone in passing into an arenaceous limestone. It may be termed the Experiment Mills shale. This shale has the stratigraphic position of and is similar lithologically to the Shriver shale. Its observed fauna is so limited that its relations are undetermined.

ORISKANY SANDSTONE: The Oriskany sandstone appears also in the northern ridge, where it is somewhat thicker and differs considerably in facies from the southern beds, being a more offshore deposit. It contains fewer pebbles, less sand, and more chert. Northeastward the pebbles disappear progressively, the sand becomes finer, and the beds more calcareous until, finally, it passes laterally into the silty and siliceous to arenaceous limestone of the Glenerie facies of Chadwick. *Spirifer arenosus* is much less frequent in this area, while *Spirifer murchisoni* is abundant.

ESOPUS SHALE: Between the Oriskany sandstone and the Onondaga limestone is an arenaceous shale, or grit, known as the Esopus shale from its typical exposure at Esopus Creek near Kingston, New York. Apart from the occurrence of the marking known as *Taonurus* its only other common species is *Leptocoelia acutiplicata*. The latter is found in abundance in some beds. More rarely undetermined species of *Chonetes* and *Spirifer* are present.

Relations of Esopus shale and Palmerton sandstone.—As long as the Palmerton sandstone was considered to be of Oriskany age the Esopus shale seemed to terminate abruptly at the southwest end of the northern ridge [41] and to have no correlate farther southwestward. The present study shows that such is not the case.

The Esopus shale and the Palmerton sandstone are both overlain by the Onondaga limestone. The Esopus shale and the Bowmanstown chert are both underlain by the Oriskany sandstone. The Palmerton-Bowmanstown beds of the southern ridge and the Esopus of the northern ridge

[41] Bradford Willard reports 87 feet of Esopus shale between the Palmerton sandstone and the Hazard paint ore in the Lehigh Valley (1939, p. 151). The authors know of no locality where this interval exceeds 10 feet; it is generally less. The beds found in this position, chiefly clays, appear to intertongue with the paint ore which is of known Onondaga age and are referred to the Onondaga.

are, therefore, of essentially the same age and replace each other in the two areas. They have also like thicknesses. The Esopus is 275 feet thick at Experiment Mills, the Palmerton-Bowmanstown beds about 260 feet thick at Little Gap, gradually thinning southwestward. The Esopus shale was deposited more off shore, while the Palmerton sandstone and Bowmanstown chert accumulated at the southeast, nearer the old shore line. Their relations are illustrated by Plate 1.

FORMATIONS OF HELDERBERG GROUP

Description of formations.—F. M. Swartz (1939, p. 24-91) has recently published a description of the Helderberg and Keyser formations of Pennsylvania to which the reader is referred for fuller treatment. Only salient facts are given here.

Port Ewen Shale: *Thlipsurella multipunctata, Homalonotus vanuxemi* zone. The Port Ewen is a calcareous shale characterized by the species named. In addition there is a considerable fauna, including *Spirifer concinnus, Chonetes aroostookensis,* several new species of *Chonetes, Dalmanites* sp., and numerous new species of Ostracoda to be described by F. M. Swartz.

Becraft Limestone: This limestone is discriminated, somewhat insecurely, in eastern Pennsylvania, chiefly because of its lithology. Farther north it is characterized by a great profusion of *Spirifer concinnus,* associated with New Scotland species.

New Scotland Formation: *Eospirifer macropleura* zone. This formation consists of calcareous shale and underlying limestone in the northern ridge, chert in the southern ridges. *Eospirifer macropleura* occurs in profusion in the limestone and shale members of the northern ridge, associated with *Spirifer perlamellosus, Spirifer cyclopterus, Coelospira concava, Meristella arcuata,* and other significant species. In the southern ridge the first-named species is rarer, but a large and diagnostic New Scotland fauna is also present there.

Coeymans Limestone: *Gypidula coeymanensis* zone. The Coeymans limestone of the northern ridge is characterized by a great profusion of *Gypidula coeymanensis. Favosites helderbergiae, Uncinulus mutabilis,* and *Strophonella punctulifera* are also common. Crinoid rings make up much of the limestone. In the northern ridge a calcareous quartz conglomerate (the Stormville conglomerate of White) bearing *Gypidula coeymanensis* (aa) is a striking feature at the top of the formation.

Correlations within the area.—The Lower Devonian beds may be followed throughout the northern ridge by their faunas and lithology. All are restricted, insofar as known, to the northern ridge, save the Oriskany and New Scotland which are found also in the southern ridge. The New

Scotland formation appears in the latter area, when weathered, as a dense, white, colloidal-looking chert, abounding, locally, in characteristic New Scotland fossils. It is overlain by the Oriskany sandstone of Ridgeley type, with the *Spirifer arenosus* fauna. No other Lower Devonian formations are known in the southern ridge save, possibly, a representative of the Coeymans which may lie in the concealed interval below the New Scotland.

FORMATIONS OF KEYSER GROUP AND KEYSER LIMESTONE

Definition of Keyser group.—The term Keyser group is proposed for the group of three formations of Keyser age—Decker sandstone, Rondout limestone, and Manlius limestone—that replace the Keyser limestone in the northern ridge. The Keyser group, as thus defined, extends from the region of study to southeastern New York.

Description of formations.—The Keyser limestone and the formations of the Keyser group are briefly characterized as follows (*see also* F. M. Swartz, 1939, p. 35-43):

Manlius Limestone: *Spirifer vanuxemi* zone. Immediately underlying the Coeymans is the Manlius limestone, characterized by *Spirifer vanuxemi* (aa), *Stropheodonta varistriata, Megambonia aviculoidea,* and *Tentaculites gyracanthus.* The fauna, though small, is distinctive.

Rondout Formation: *Leperditia* sp. zone, in part. This formation is twofold. The lower part is a laminated limestone bearing a great profusion of *Leperditia* sp. This is the Decker's Ferry limestone of I. C. White (1882, p. 137). The upper part is a barren, argillaceous limestone, becoming, when weathered, "pethstone-like." It is the Stormville hydraulic cement bed of I. C. White (1882, p. 136).

Decker Sandstone: *Chonetes jerseyensis* zone. The Decker sandstone, forming the lowest member of the Keyser group in the northern ridge, is characterized by the profuse occurrence of *Chonetes jerseyensis,* associated with *Camerotoechia litchfieldensis, Stenochisma deckerensis, Stropheodonta (Leptostrophia) bipartita* and other distinctive forms.

Keyser Limestone: *Chonetes jerseyensis* zone. The three formations named above are found in characteristic development in the northern (Godfrey-Wallpack) ridge. In the southern ridge they lose their identity. The lower beds are replaced by an impure, argillaceous, and arenaceous limestone which is overlain, locally, by argillaceous sandstones and shales. The limestone is characterized by the presence of *Chonetes jerseyensis* (aa), *Rhynchospira globosa* (a), *Stropheodonta (Leptostrophia) bipartita,* and numerous Ostracoda, including *Kloedonella ventricornis* var. and *Zygobeyrichia barretti.*

The fossiliferous beds are overlain at Andreas and Snyders by red-brown sandy beds of Bloomsburg type, which may, perhaps, represent some of the upper Keyser beds of the northern ridge, or even, possibly, the Coeymans. No fossil evidence of their age is known.

Correlations within the area.—The Manlius, Rondout, and Decker formations may be traced, in characteristic development, throughout the entire northern ridge. In passing into the southern ridge the above formations are replaced by the Keyser limestone, with the *Chonetes jerseyensis* fauna in its lower part. Beds holding the latter fauna can be traced from southeastern New York to the western limit of the limestone exposures in the area under consideration.

SILURIAN FORMATIONS

Description of formations.—The Keyser limestone and its correlates are underlain in the area by the following formations, in descending order: Bossardville limestone, Poxono Island shale, Bloomsburg red beds.

BOSSARDVILLE LIMESTONE: The Bossardville limestone, lying immediately below the Keyser group throughout its extent from southeastern New York to Schuylkill Haven, Pennsylvania, is the uppermost formation of undoubted Silurian age in the area. When typically developed, as in the vicinity of the Delaware Water Gap, it consists of three parts. Its major (central) beds are a blue-black, dense limestone, quarried extensively because of its purity and adaptation to burning. Some of these beds, show, generally, the laminated structure so characteristic of the Tonoloway limestone farther southwest. Other beds, however, do not exhibit this feature. At many places the limestone is riven by cracks, since filled by light-colored calcite, presenting a striking appearance. This quarry division is 50 or 60 feet thick in the type area. Underlying these beds, at some places, are thin-bedded, gray, finely straticulate and somewhat shaly limestones interrupted by mud cracks, which under pressure have developed prismlike pillars, described by I. C. White (1882, p. 144). Overlying the quarry rock, locally, is a few feet of thick-bedded, impure limestone weathering yellowish. White (1882, p. 141) describes green shales, the "Decker's Ferry shale" which the present writers have not seen, as lying at the top of the formation.

Leperditia sp. is common to abundant in some of the beds, especially in the lower division in the northern ridge and the eastern part of the southern ridge. They have not been specifically identified, and no other recognizable species has been seen in that area. Other species of Ostracoda make their appearance farther southwest and become numerous at the Schuylkill River. (For discussion of latter see description of sections at Schuylkill Haven.)

Poxono Island Shale: The Poxono Island shale underlies the Bossardville limestone. It consists of calcareous shale weathering light-colored, greenish to buff. Interbedded with the shale are local red bands of Bloomsburg type. Near the base of the shale, at the type locality, is a bed of blue-gray limestone, about 5 feet thick—White's (1882, p. 147) Poxono Island limestone—containing great numbers of Ostracoda too poorly preserved to permit confident identification. Other than these no fossils have been observed in this formation.

The thickness of the shale is subject to great variation. At the Schuylkill River it is absent or reduced to a thickness of a few feet. It gradually thickens northeastward. At Snyders it is about 150 feet thick. At the Lehigh River the thickness could not be determined, but thin red tongues appear in the shale 200 feet beneath its top. At the Delaware Water Gap the shale is about 200 feet thick. White gives the same thickness at the type locality. It seems to diminish in thickness northeastward.

Bloomsburg Red Sandstone: Underlying the Poxono Island shale, and intertonguing with it, at places, is the Bloomsburg red sandstone and shale. It consists of characteristic redbeds, in which are occasional bands of greenish or lighter color. Two localities give good measurements of thickness—Rauschs, southwest of New Ringold, where it is 1815 feet thick, and Fox Gap, near Delaware Water Gap, where it is 1900 feet thick. It is, perhaps, 2000 feet thick at Lehigh Gap.

The Bloomsburg sandstone is manifestly a facies of varying age as shown by the authors (1931, p. 656-658) in an earlier publication.

Underlying Formations: The Bloomsburg is underlain, in this area, by the Clinton shale and Tuscarora sandstone at the Schuylkill River and by the Shawangunk conglomerate, of combined Clinton and Tuscarora age, at the Delaware River. These formations have been discussed elsewhere (Swartz and Swartz, 1931, p. 639-661).

Correlations within the area.—No identifiable fossils have been observed in the Poxono Island shale and Bloomsburg redbeds, and only *Leperditia* sp. is common in the Bossardville limestone at most localities, so that faunas afford little aid in the correlation of these formations in the area under consideration. The stratigraphic sequence, distribution, and lithology of the formations in closely placed sections, however, permit confident correlations. Above the series is the Keyser group with the diagnostic *Chonetes jerseyensis* fauna. The Bossardville limestone can be traced beneath the Keyser group throughout the entire area, becoming more shaly and impure toward the southwest. The Poxono Island shale vanishes toward the southwest where it is probably replaced, gradu-

ally, by intertonguing Bloomsburg redbeds. The Bloomsburg can be traced by its lithology and characteristic soil throughout the area. Beneath the Bloomsburg are the beds previously described by the authors.

CORRELATIONS WITH FORMATIONS OF OTHER AREAS

Correlations with formations of northern New Jersey and southeastern New York.—The sequence described above can be traced from Decker's Ferry to northern New Jersey, where it was described by Weller, and thence into southeastern New York, where it was studied by Shimer. The distance is so short, 24 miles, the exposures so close, work so detailed, and faunas so large as to permit little doubt of the correlations made by the various workers in that area.

Correlations with formations of central Pennsylvania, Maryland, and West Virginia.—The Palmerton sandstone and Bowmanstown chert do not extend beyond the region under consideration but are represented southwestward by a well-defined unconformity between the Oriskany sandstone and Onondaga (Romney) shale or limestone. The Lower Devonian and Keyser formations are well developed in central Pennsylvania where they occur in normal sequence and with characteristic faunas. They have been so fully described by different workers [42] that no further discussion of them is needed here. The Bloomsburg red beds can be followed continuously from southeastern New York and northeastern Pennsylvania to central Pennsylvania and Maryland. Only the Bossardville limestone and Poxono Island shale, therefore, need consideration here. Relations of the latter formations are less manifest.

The Bossardville limestone and Poxono Island shale of eastern Pennsylvania are comparable in lithology and stratigraphic position to the Tonoloway limestone and Wills Creek shale, respectively, of central Pennsylvania and Maryland. Both the Bossardville and Tonoloway limestones are immediately overlain by beds of early Keyser age, containing the large and diagnostic fauna of the *Chonetes jerseyensis* zone of the Keyser. Both the Poxono Island and Wills Creek shales are directly underlain by the Bloomsburg red beds which are continuous from one area to the other. Their thickness, however, differs, the thickness of the combined Wills Creek and Tonoloway formations being much greater than that of the Poxono Island and Bossardville.

In spite of their similarity the Poxono Island and Bossardville formations and the Wills Creek and Tonoloway beds of Maryland and central Pennsylvania are believed to have different time limits. The evidence for this is as follows:

[42] Lower Devonian, Maryland Geol. Surv. (1913), J. B. Reeside, Jr. (1918), A. B. Cleaves (1939), F. M. Swartz (1939).

TABLE 1.—*Early Devonian and Late Silurian formations of Eastern Pennsylvania*

(Dashes denote absence)

		Auburn	Schuylkill Haven	South Ridge Stony-Chestnut Ridge	North Ridge Godfrey-Wallpack Ridge
Devonian	Middle	Onondaga limestone -	Onondaga limestone -	Onondaga limestone Palmerton sandstone Bowmanstown chert	Onondaga limestone Esopus shale
	Lower	Oriskany sandstone -	Oriskany sandstone -	Oriskany sandstone - New Scotland chert Coeymans (?) limestone	Oriskany sandstone–ls. Oriskany shale Port Ewen shale Becraft limestone New Scotland shale New Scotland cherty ls. Coeymans limestone
Silurian or Devonian		- -	Keyser formation (*Chonetes jerseyensis*)	Keyser formation (*Chonetes jerseyensis*	Manlius limestone Rondout limestone Decker sandstone (*Chonetes jerseyensis*)
Late Silurian		- Bloomsburg sandstone	Bossardville limestone - - - - - - ? - - - - - - Bloomsburg sandstone	Bossardville limestone Poxono Island shale Bloomsburg sandstone	Bossardville limestone Poxono Island shale Bloomsburg sandstone

The Bloomsburg occurs at the southwest as a thin red bed at, or near, the base of the Wills Creek. It increases in thickness northeastward, as shown in many sections, by the successive appearance of higher and higher red tongues above the lower beds and their progressive thickening northeastward. Thus at Keyser, West Virginia, the red bed is 2 feet thick, at the base of the Wills Creek; near Hancock, Maryland, red bands appear in the lower 100 feet of the Wills Creek; at Fairview Mountain, west of Hagerstown, Maryland, the Bloomsburg is 200 feet thick, and red bands range upward, through the Wills Creek, into the lower Tonoloway. At the same time red beds appear in the midst of the underlying McKenzie. At Loysville, Pennsylvania, northwest of Harrisburg, the Bloomsburg becomes 500 feet thick, and the red bands, in the overlying Wills Creek and Tonoloway, have increased in number and thickness. In this process of replacement some of the red tongues can be traced from section to section.

Finally, at Swatara Gap, 24 miles northeast of Harrisburg, the Wills Creek shale and Tonoloway limestone have wholly vanished (the upper part of the latter, perhaps, by unconformity), the Bloomsburg is over 1500 feet thick, and red beds replace the upper part of the underlying McKenzie. Farther eastward the Bloomsburg increases in thickness to 2000 feet at Lehigh Gap, and the Poxono Island shale and Bossardville limestone appear above the Bloomsburg and below the Keyser fauna.

These facts seem to show that both the Poxono Island shale and Bossardville limestone are later than the Wills Creek shale and, perhaps, later than the earlier Tonoloway. According to this view the Wills Creek shale and Tonoloway limestone of Maryland and central Pennsylvania and the Poxono Island shale and Bossardville limestone of eastern Pennsylvania are similar and homotaxial, but not synchronous pairs, the Poxono Island being younger than the Wills Creek, while the Bossardville represents the upper part of the Tonoloway.

These relations are illustrated in a diagram showing numerous sections in an earlier publication of the authors (1931, p. 660, Fig. 2).

DISTRIBUTION AND GEOGRAPHIC VARIATION OF FORMATIONS

Table 1 and the chart of Columnar sections (Pl. 1) show clearly the distribution and variation in character and thickness of the sediments. The Lower Devonian sediments are best developed in the northern ridge; of these, the Oriskany sandstone and the New Scotland beds are the only formations certainly known to pass from the northern into the southern ridge, although the Coeymans may also do so.

All the Devonian formations show an increasingly arenaceous content toward the southeast, manifestly due to approaching the old shore line in that direction and the overlap of formations upon the shore. In going southwestward the Middle Devonian Esopus passes into the conglomeratic Palmerton sandstone and Bowmanstown chert, the Oriskany loses its basal shale member and assumes the facies of the Ridgeley sandstone with *Spirifer arenosus*, and the New Scotland loses its shale member and becomes a chert (or cherty limestone?). Only the lower Keyser becomes more calcareous, though remaining arenaceous.

Farther southwest the formations gradually thin and disappear, until at Schuylkill Haven only the Bossardville, Keyser, and a few feet of Oriskany sandstone persist. Six miles away, at Auburn, all the formations between the Bloomsburg and Onondaga have vanished, save a few inches of questionable Oriskany.

AGE OF THE KEYSER LIMESTONE

That the Keyser limestone is the correlate of the Manlius, Rondout, and Decker formations, inclusive, of northern New Jersey and southeastern New York is accepted by all students of the Keyser, however diverse their views as to the age of the latter may be. Maynard (1913, p. 97) who studied the Keyser in Maryland, Reeside (1918) who investigated it in central Pennsylvania, Ulrich (1912) to whom we owe its name, and the writers (C. K. Swartz, 1913, p. 110-117; F. M. Swartz, 1939, p. 43-46) all agree in this respect.

That the Keyser is older than the typical Helderberg is evident, since it is unconformably overlain by the Coeymans throughout much of its extent. The question is not, therefore, as to the relation of the Keyser to the Manlius and older formations of southeastern New York but, first, as to its relation to the typical Manlius of central New York and, secondly, as to the relation of the latter to the European Silurian.

That the Manlius of southeastern New York is the correlate of the Manlius of central New York seems very probable. Ulrich believed this not to be true (C. K. Swartz, 1913, p. 115-117), but more recent investigations lend weight to the view, long held by the geologists of New York, that the Manlius of both areas is of essentially the same age. If this view is correct the Keyser is not younger than the typical Manlius of central New York.

That the Helderberg of eastern New York is Lower Devonian was shown to be true, with a high degree of probability, by J. M. Clarke in 1889 and by Charles Schuchert in 1900. So strong was the evidence

submitted that nearly all students accepted their conclusions, though with some notable exceptions.

The view, however, that the Manlius of New York is uppermost Silurian does not seem to rest upon equally assured evidence. The Helderberg is an open-sea formation with abundant life. The Manlius of central New York was deposited in a highly restricted sea with a restricted fauna. The great dissimilarity of their faunas seems to be an expression of their ecology, rather than a trustworthy indication of time relations.

In northern New Jersey and southeastern New York the sea was manifestly somewhat restricted during Keyser time. Weller (1903, p. 72-80) saw in its fauna nothing to indicate Devonian relations and did not hesitate to make its deposits Silurian.

In passing from New Jersey to Maryland and West Virginia we come into a more open sea, the formations become thicker, more calcareous, the faunas more normally marine. With these conditions appear a number of incoming Devonian species. A few of the latter make their first appearance in the lower part of the Keyser. More appear toward its upper limit (Swartz, C. K., 1913, p. 118-120). Ulrich, impressed by these facts, and attaching more weight to the incoming members of the fauna than to the older survivors, made the Keyser Lower Devonian and assigned it to the Helderbergian, to which its new forms are clearly related. The larger number of Keyser species, however, have Silurian affinities, while a smaller number are Devonian types.

Whether we make the Keyser Silurian or Devonian will depend upon our guiding principles. If the first appearance be chiefly weighed, as done by Ulrich,[43] it becomes Devonian. If equal or nearly equal weight be attached to both types, it is Silurian.

It is clear, however, that a comparison between the European and the American faunas must be made using the fauna of the Keyser as a basis of comparison and not that of the Manlius, with its restricted conditions of life. A fundamental study of this problem remains to be undertaken. It is only as we compare reasonably similar marine horizons that trustworthy results can be secured.[44]

[43] The writers question attaching as much weight to first appearances as is done by Ulrich.

[44] Relations of faunas of the uppermost Silurian and early Devonian have recently been reviewed by R. S. Allan [The Fauna of the Reefton Beds, New Zealand, New Zealand Geol. Surv., Palaeontological Bull. No. 14], (1935, p. 46-59), chiefly on the basis of an extended review of the existing literature and examination of museum collections of European fossils. Allan correlates the Helderberg with the European Siegenian and the Tonoloway with the Lower Gedinnian by their faunas. He correlates the Keyser with the nonmarine Upper Gedinnian beds of western Europe by its position but believes that the marine Keyser fauna has no representative in western Europe. Allan proposes to make both the Gedinnian and Keyser Silurian. However, original workers in western Europe, such as Maillieux and Asselberghs, have in general continued to include the Gedinnian in the Devonian system.

(Continued bottom next page.)

Rather than set aside the generally accepted classification of the Manlius, and in view of the uncertainties in correlation of the Keyser and Manlius with European horizons, the writers have placed the Keyser in the Silurian system, but with a question mark, as the true status of the Keyser is yet to be determined.

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In his classic work on the age of the Helderberg faunas, Schuchert regarded the Siegenian as corresponding to the Oriskany rather than to the Helderberg of North America. In any event, the "*Spirifer hystericus* fauna" of the Siegenian contains such important post-Helderberg types as *Eodevonaria, Meganteris, Athyris, Tropidoleptus,* the multiplicate *Spirifer pellico,* and large-muscled *Stropheodonta gigas.*

Evidence for the correlation of the Tonoloway and early Gedinnian is still less satisfactory. It is clear that the age relations of the Keyser with European horizons is not yet firmly established and that solution of the problem must await further investigations.

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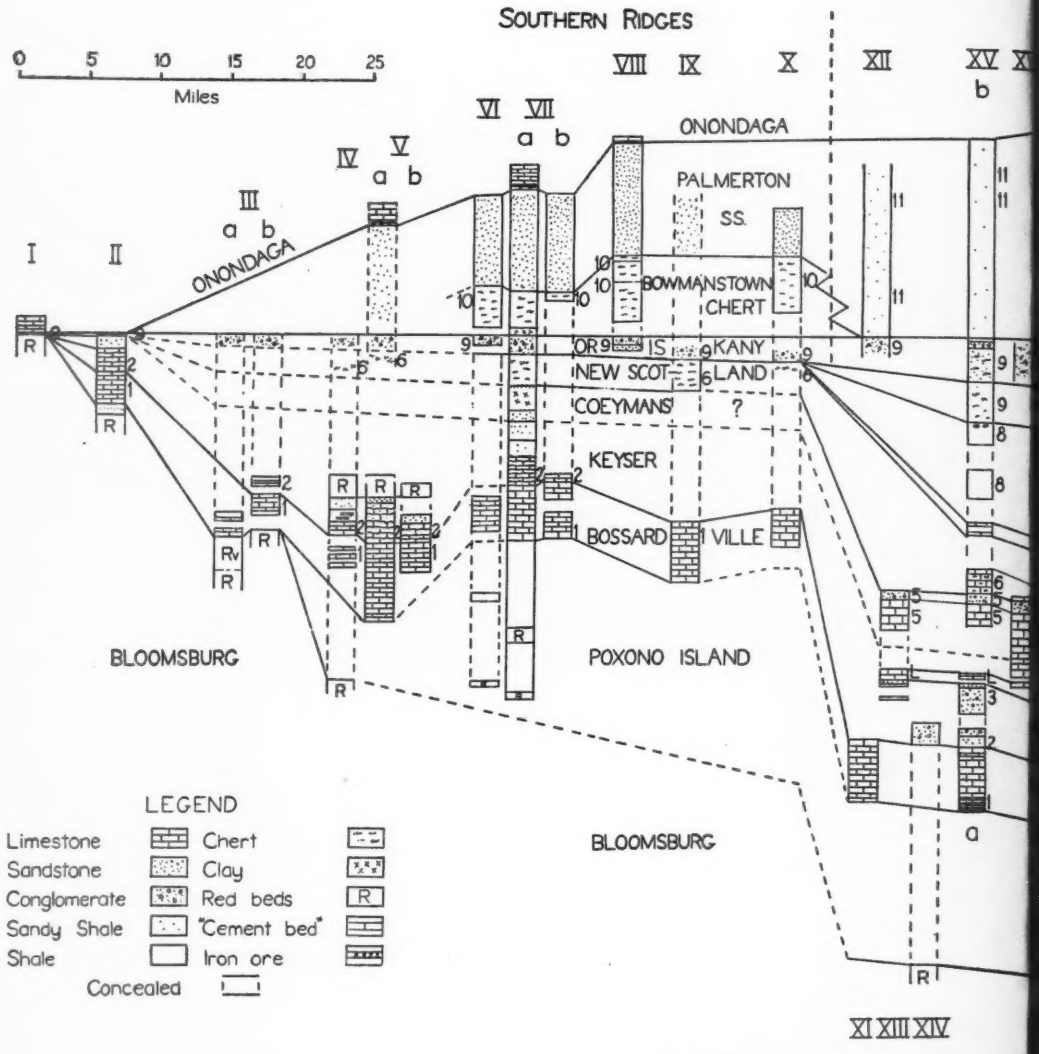
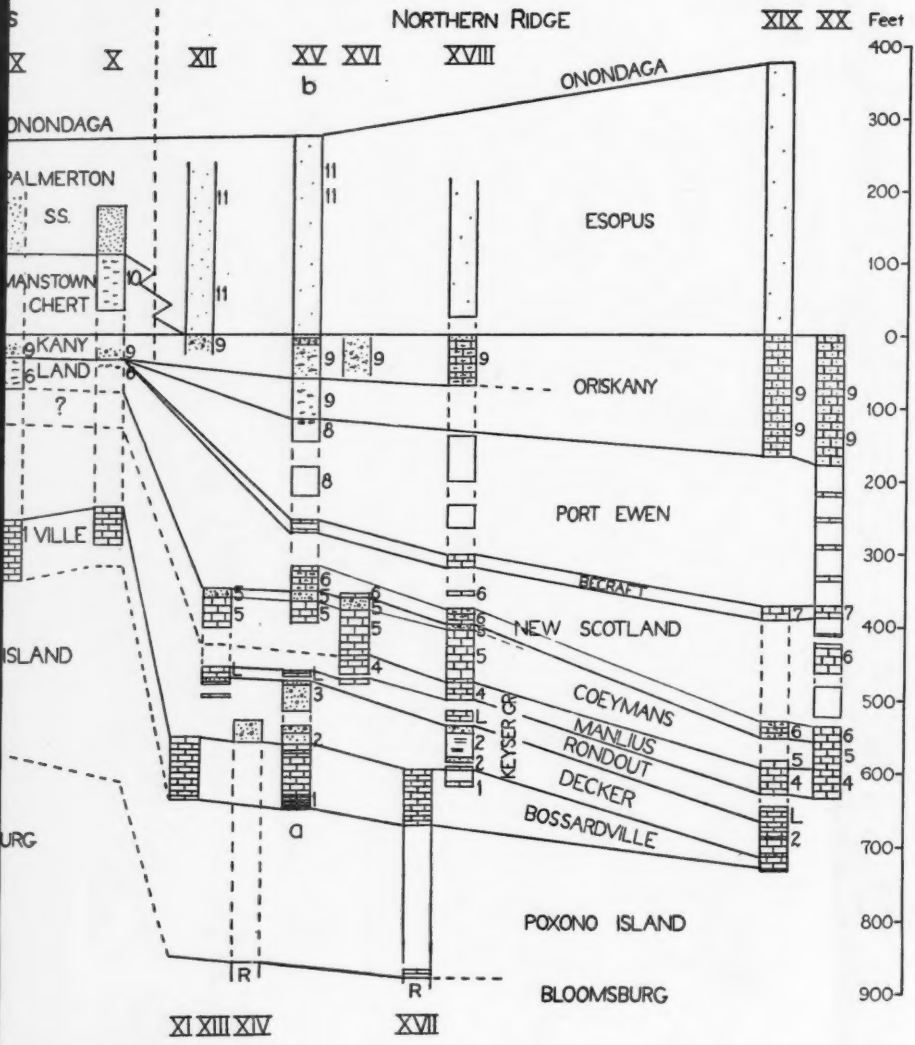


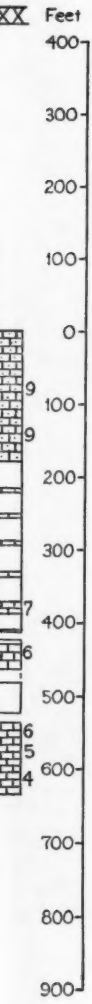
COLUMNAR SECTIONS

Sections represented by vertical columns, drawn to scale. Sides, interrupted lines where beds are conc
Figure 1. Arabic numbers signify faunas. All sections are by the authors unle



COLUMNAR SECTIONS

es, interrupted lines where beds are concealed. Locations shown by Roman numerals in unas. All sections are by the authors unless otherwise stated.



SECTIONS

Pennsylvania
Schuylkill County
I Auburn
II Schuylkill Haven
III New Ringold
a. Southwest of
b. East of
IV Snyders
V Andreas
a. Quarry J. Rudolph
b. Quarry Wm. Loch
Carbon County
VI Bowmanstown
VII Palmerton
a. Ore tunnel near Hazard
b. Quarries
VIII Little Gap
Monroe County
IX Kunkletown
X Saylorsburg
XI Bossardville
XII Godfrey Ridge
XIII Stormville
XIV Fox Gap
XV Delaware Water Gap
a. Croasdale Manor
b. Experiment Mills
XVI Shawnee-on-the-Delaware
XVII Poxono Island, by I. C. White
XVIII Decker's Ferry
New Jersey, Sussex County
XIX Nearpass quarry, by S. Weller
New York, Orange County
XX Trilobite Mountain, by H. W. Shimer

FAUNAS

1. (L.) *Leperditia* sp. fauna
2. *Chonetes jerseyensis* fauna
3. *Spirifer modestus* fauna
4. *Spirifer vanuxemi* fauna
5. *Gypidula coeymanensis* fauna
6. *Eospirifer macropleura* fauna
7. *Spirifer concinnus* fauna
8. *Thlipsurella multipunctata-Homalonotus vanuxemi* fauna
9. *Spirifer arenosus-S. murchisoni* fauna
10. *Spirifer macrus* var. fauna
11. *Leptocoelia acutiplicata* fauna.

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MULTIPLE GLACIATION OF MAUNA KEA, HAWAII

BY CHESTER K. WENTWORTH AND WILLIAM E. POWERS

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ABSTRACT

Mauna Kea, Hawaii (13,784 feet), was glaciated four times during a period presumably correlative with the ice age elsewhere. Snow mantles Mauna Kea during winter but banks rarely survive the summer. Evidence is abundant not only for the latest glacial stage in the summit area but also for three earlier glacial advances in the zone outside and below the youngest moraine. The four indicated stages have been named, beginning with the latest, the Makanaka, Waihu, Pohakuloa, and pre-Pohakuloa stages.

Distinguishing features of the several drifts are:

(1) Stratigraphic position. The three earlier ones lie under successive series of late lava flows.
(2) Moderate weathering and surface staining (to brown) of the older drifts. Climate has probably been periglacial throughout Pleistocene and recent time.
(3) Lithologic differences due to derivation from different series of surface lavas.
(4) Matrices of the older drifts are partly tuffaceous indicating contemporary volcanism.
(5) In places the oldest drifts are well-indurated tillite.
(6) The drifts are dominantly boulder beds, much water-washed. Boulders are somewhat faceted but only faintly striated.
(7) Striated pavements are known under the older drifts, but glacial erosion was generally feeble.
(8) Glaciers in the pre-Pohakuloa stage descended to about 7000 feet but only to 10,200 feet in the latest stage.
(9) Interglacial processes other than extrusive vulcanism are little known. Climate is now subarctic above the timber line; and significant soils probably were never developed.

INTRODUCTION

STATUS OF GLACIAL STUDIES

Although the summit of Mauna Kea was first visited by a European in 1823, and was subsequently studied and described by numerous observers of note including the geologists C. E. Dutton and C. H. Hitchcock, the former glaciation of this mountain was not recognized until 1910 (Daly, 1910, p. 297). Since that time the boulder moraines, striated ledges, and other unmistakable glacial features of the summit area have been observed by numerous persons including geologists, topographers, and foresters. The first systematic account of these glacial features was presented in 1937, following an expedition in 1935 by the Hawaiian Academy of Science but based in part on earlier studies by Gregory (Gregory and Wentworth, 1937). That paper gives a brief account of exploration of the mountain. Only the higher portions of Mauna Kea were mapped by Gregory and Wentworth, and their descriptions pertain almost wholly to the results of the latest glacial stage, although some evidences of earlier glacial stages were noted.

THE PRESENT INVESTIGATION

To pursue the study of earlier drifts and multiple glaciation, the authors spent 5 weeks on Mauna Kea in the summer of 1939. Although further details of the latest glaciation were mapped in the summit area lying mainly above 11,000 feet, most attention was given to the zone between 8000 and 11,000 feet on the south side. Here are abundant exposures of the older drifts interstratified with lava flows and bearing various evidences of antiquity. Here also were found two of the three deepest gulches on Mauna Kea, namely, Waikahalulu and Pohakuloa. Although only a few hundred feet deep, both exhibited excellent sections of the older drifts and interstratified lavas. A base camp was established on the south side of the mountain at Hale Pohaku, near the timber line at about the 9300-foot level. From here the south, southwest, and southeast slopes below 12,000 feet were carefully studied by foot traverse, and all drift areas were mapped. By use of pack train and temporary camps, the less accessible northern, northwestern, and northeastern slopes were examined though less carefully than the south. These studies indicated that the best exposures and clearest relationships between the several drifts are to be found on the south and fortunately more accessible side of the mountain. However, further study is desirable in the sector northwest of the summit and in the eastern sector between Waipahoehoe Gulch and Puu Makanaka.

ACKNOWLEDGMENTS

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PHYSIOGRAPHY

LOCATION

Mauna Kea, highest point in the entire Pacific, forms the north-central part of the island of Hawaii (Fig. 1). The summit at 13,784 feet is slightly higher than Mauna Loa (13,680 feet) lying 25 miles to the south, although the latter is much broader and more massive. These and three

smaller volcanic domes—Kilauea, Hualalai, and Kohala—form a rudely triangular island approximately 90 miles on a side. The considerable area of the island (4030 square miles) and the great height of Mauna Kea

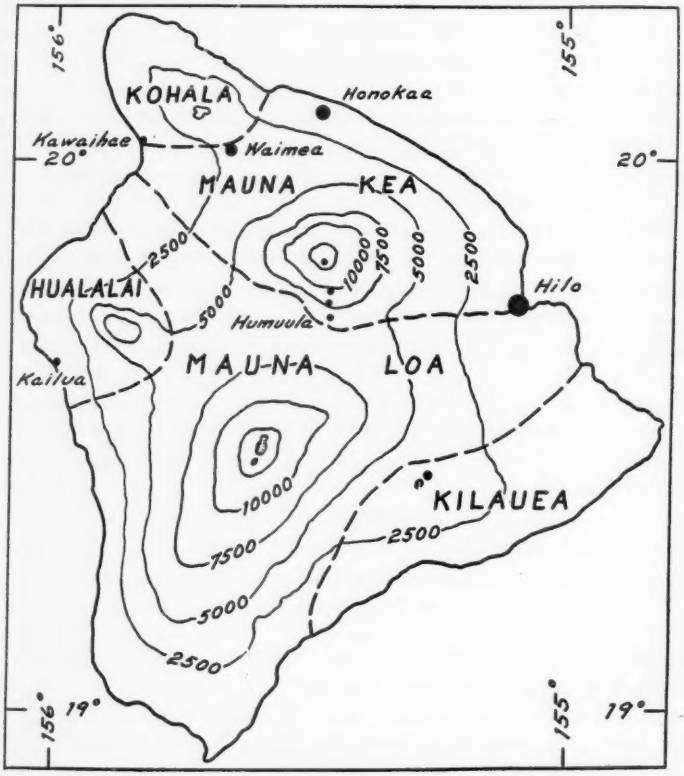

FIGURE 1.—*Sketch map of Island of Hawaii*
The five domes are shown by contours, interval 2500 feet.

and Mauna Loa exert a strong influence on its present climate and undoubtedly were major factors in the formation and long duration of Pleistocene ice caps. Mauna Kea, Hualalai, and Kohala reached their present size before or during the ice age and have suffered little or no volcanic upbuilding since. Mauna Loa is still active and computation based on present growth rate indicates that it must have been only 11,000 to 12,000 feet high during the latter part of the glacial epoch. Topography of the

island controls the present distribution of rain and snow and must be considered in an analysis of conditions leading to glaciation in the past.

TOPOGRAPHY

Mauna Kea is a great lava pile rudely conical but characterized by marked steepening in slope in the zone approximately between 7000 and 11,500 feet. Above this zone the summit area flattens into what may be called the summit plateau. Below 7000 feet slopes also decrease giving the volcano a broad massive base. The lower slopes continue to the sea on the northeast and northwest, but on the south meet Mauna Loa in a saddle at about 6600 feet (Fig. 1), where Mauna Loa has grown against Mauna Kea. For the details of topography the reader is referred to the Mauna Kea, Waikii, Humuula, and Kaohe topographic quadrangles of the United States Geological Survey. These maps show accurately all parts of the mountain above approximately 3500 feet, and the first two include the summit area down to approximately 8500 feet on the south and 3500 feet on the north.

The form of Mauna Kea had a marked effect on the nature, movements, and extent of glacial ice and hence deserves further description. The lower mountainside below the 7000-foot contour slopes 300 to 500 feet per mile, a value controlled by flow conditions of lava during the major period of growth of the volcano. In the steepened zone above 7000 feet, it slopes at a rate of 850 to 2000 feet per mile. Interbedding of flows with beds of fragmental ejecta is widespread in the upper zone, and a change in type of activity from the earlier liquid phase is reflected by the steeper slopes here. Many of the lava flows in this zone have slopes of 20 degrees to 24 degrees. Above 11,000 to 12,000 feet is the summit plateau, a rudely circular dome 5 or 6 miles in diameter rising between 500 and 1000 feet per mile to a central area above 13,000 feet. The actual summit at 13,784 feet is a cinder cone, similar to those which are widely distributed over the summit area and sides of Mauna Kea. Why the summit area flattens is unknown, but it probably has some connection with volcanic growth rather than glacial erosion. The latter, although confined largely to the summit area, was relatively feeble.

More than 80 cinder cones have been built on the summit plateau and sides of Mauna Kea, with a marked concentration in three zones extending radially northeast, south, and west from the summit. Many are only a few hundred feet in diameter; largest is Puu Makanaka in the northeast zone, nearly a mile wide and 600 feet high, with a crater 1600 feet across at the rim (Fig. 1; Mauna Kea quadrangle map). On the surface the majority of these cones show only cinders 8 to 12 millimeters in size but some are strewn with coarser driblet lava and bombs. These cones

developed during an explosive closing phase of Mauna Kea's upgrowth, almost wholly before the last stage of glaciation, and their relations to the latter process are made clear by the distribution of blocky moraine. Many of the smaller cones were completely covered by the glacier, but the larger ones are block-strewn only about their lower portions and thus serve as measures of the thickness of the surrounding glacial ice. Detailed descriptions of many of these cones have been presented elsewhere by Wentworth (1938, p. 59-72).

A factor of importance in the accumulation and movement of glacial ice on Mauna Kea is the complete absence of sizable valleys. In the summit plateau and steepened outer zone, stream channels are few and small, and practically no permanent streams exist. Despite the fact that the average slope approaches 40 per cent in many places, erosion by running water has been feeble in consequence of the extremely porous character of the cinder and lava surfaces and the relatively light precipitation occurring at the higher levels. Stream channels more than 25 to 50 feet deep, or wider than small ravines, are uncommon. The deepest of these radial valleys are Pohakuloa, Waikahalulu, and Hanipoe gulches on the southwest, south, and north, respectively. Both of the former gulches are from 100 to 300 feet deep for distances of approximately 2 miles in the zone between 8000 and 11,000 feet. Hanipoe Gulch is somewhat more shallow. Slight as these depths are, they encouraged ice advance so that well-formed lateral boulder moraines of the last stage flank the sides of Hanipoe and Pohakuloa Gulches for half to three quarters of a mile beyond the main area of marginal moraine.

Average slopes are so steep in the zone just outside of the summit plateau that the absence of deep radial valleys can be attributed only to deficient runoff. Loose cinder slopes and highly porous lava flows favor infiltration of most rainfall, and the water table lies below the shallow channels. Only one flowing stream was observed above 7000 feet by the writers in July, 1939, namely, the west branch of Pohakuloa Gulch for a short distance above the 10,000-foot level. Here a water table, very likely of the perched type, is held up by a bed of early glacial drift and gives rise to a flow of small volume where the channel is cut through overlying lavas. This stream dwindles and disappears at about the 10,000-foot level, less than half a mile from its source. Half a dozen other springs and seepages were discovered, all apparently caused by compact beds of the older drift lying below volcanics, but in no case giving rise to channel flow. Nevertheless, abundant freshly scoured potholes and rock channels are found throughout most of the gulches (Pl. 3, fig. 1), indicating effective stream flow and scour presumably in spring when the snowcap of the summit area melts. The character of these

channels throws light on the former glaciation of Mauna Kea. Stream scour, effective even now, must have been much greater during melting of the icecap when stream flow was able to carry boulders up to 6 feet

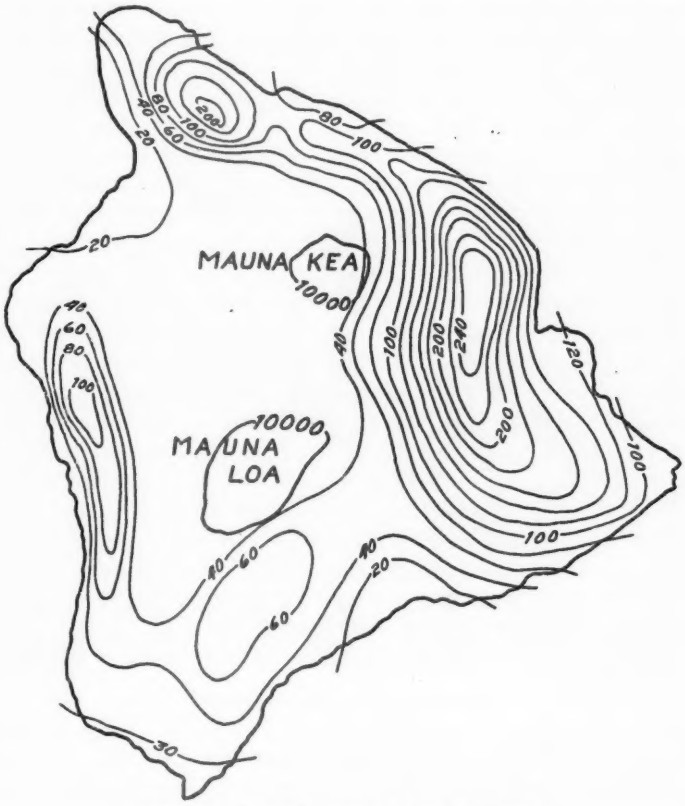

FIGURE 2.—*Isohyetal map of Island of Hawaii*
Mean annual rainfall in inches. (Data from Territorial Planning Board.)

in size far beyond the ice margin. Further, freshly stream-scoured surfaces in some ways bear considerable resemblance to the polished and striated surfaces produced by glacial erosion. Lateral shifting and incision of the stream channel have left numerous surfaces of this type well above and somewhat distant from the present position of stream flow. Confusion with a glacier pavement is possible, particularly where an accumulation of large erratic boulders is present. Only where the striated

surface is well removed from positions of probable stream flow or is associated with indisputable beds of ice-laid till, can one conclude with confidence that the area under consideration was actually glaciated.

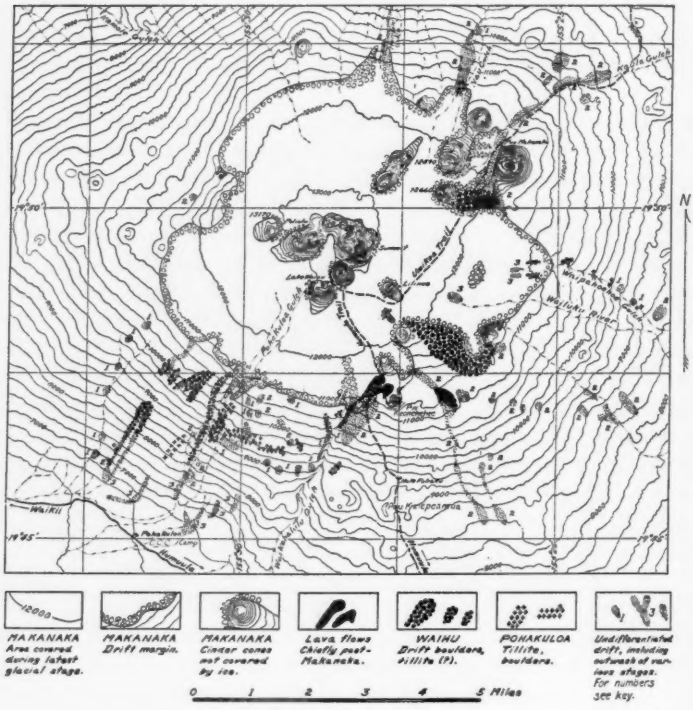
N
MAKANAKA Area covered during latest glacial stage.
MAKANAKA Drift margin.
MAKANAKA Cinder cones not covered by ice.
Lava flows Chiefly post-Makanaka.
WAIHU Drift boulders, tillite (?).
PONAHULOA Tillite, boulders.
Undifferentiated drift, embodying outwash of various stages. For numbers see key.
0 1 2 3 4 5 Miles
CONTOUR INTERVAL : 250 FEET.

FIGURE 3.—*Map of glacial features of Mauna Kea*

Significance of numbers used to designate undifferentiated drift patches: (1) Close-spaced and (2) scattered erratic boulders, transporting agent in doubt; (3) Indurated drift.

DRAINAGE

The drainage pattern on Mauna Kea is radial and except at low levels, poorly developed. Figure 3 shows that no permanent and only a few intermittent streams originate on the higher parts of the mountain. A better picture of the drainage pattern is gained from the United States Geological Survey's topographic sheets (*see* Mauna Kea, Waikii, Waipio, Hamakua, Honomu, Hilo, Humuula, and Kaohe quadrangles) which indicate that continuous intermittent streams number only 5 at the 12,000-

foot level, 22 at 10,000 feet, and 91 at 6500 feet, while permanent and intermittent streams together number 223 at the 3500-foot level. The number and depth of channels at various elevations reflect differences in rainfall, in character of soil and subsurface rock, and in the erosion period subsequent to volcanic or glacial interference. The great increase in number and flow of streams below 3500 feet is due to the heavy rainfall which probably reaches a maximum between 2000 and 3000 feet (Feldwisch, 1939). Weathering has been very much more rapid at the lower and moister levels, and its products have decreased porosity of surface materials and increased runoff. The sparseness of streams and stream channels in the summit area indicates the relatively light precipitation and ineffective chemical weathering processes at these levels as well as the relatively short period since glaciation. At any given level the number and spacing of streams differs considerably in different sectors, as shown by the United States Geological Survey topographic sheets. Field study shows that areas of relatively fresh lava flows contain no streams and that many areas devoid of streams bear evidence of volcanic activity later than that in the adjacent sectors. Distribution of streams in the summit area indicates that precipitation bears little relation to exposure to the prevailing wind. Precipitation appears to be rather uniformly distributed at high levels and is due more to high-level storms from the southwest (or Kona) direction than to orographic lifting of the trade winds that blow prevailingly from the northeast below 10,000 feet. The most interesting drainage feature of Mauna Kea is Lake Waiau, located in the breached crater of a cinder cone at 13,007 feet, and draining into Pohakuloa Gulch. A description of this water body has been given by Gregory and Wentworth (1937, p. 1726-1727).

CLIMATE

That an oceanic island lying within 20 degrees of the equator could be glaciated down to the 7000-foot level is extraordinary, and its implications regarding Pleistocene world climates are important. Glacial climate on Mauna Kea can be correctly evaluated only by first understanding present conditions. Therefore previous observations will be summarized briefly and supplemented by data from the present expedition (Coulter, 1938; Raine, 1939; Wentworth, 1939a; Gregory and Wentworth, 1937).

More than 240 inches of rainfall occurs annually on the east slopes of Mauna Kea at elevations between 2000 and 3000 feet (Fig. 2; Feldwisch, 1939, Pls. 51, 56). Eastward (*i. e.*, downward) from this zone the amount diminishes to 140 inches at Hilo which is at sea level, while westward the amount also diminishes to 40 inches at 10,000 feet. The summit area,

western slope, and all but lower part of northern slope receive less than 40 inches. It seems unlikely that the summit area near Lake Waiau receives more than 15 inches including both snow and rain.

Snow is common above 10,000 feet during winter months and occasionally occurs in summer. During winter storms a light snow cover frequently mantles the mountain above 8000 to 9000 feet and once, in 1936, extended down to 7000 feet. Deep drifts form in the summit area, and sometimes small snow patches endure throughout the summer. During early August, 1935, no snow was observed but in late July, 1939, residual snowbanks up to 4 feet in depth were present, with well-developed sun cups (Pl. 3, fig. 2; Wentworth, 1939b; 1940).

TABLE 1.—*Temperature (in °F.) for lower levels near Mauna Kea*

(Data from United States Weather Bureau, Bull. W, 1926.)

Station	Hilo (sea level)	Waimea (2700 ft.)	Humuula (6685 ft.)
Mean annual	72.1	62.9	52.0
Mean January	69.7	61.1	49.8
Mean July	73.9	64.0	54.6
Lowest	51	34	25
Highest	91	83	84
Mean range during January*	15.3	14.3	21.9
Mean range during July*	14.1	13.8	20.4

* Computed from mean maximum and mean minimum for these months.

No continuous long-term temperature records are available for the summit area, but an approximation of the summit temperatures may be reached by extrapolation from data for lower levels near Mauna Kea, as indicated in Table 1.

Extrapolation from these figures indicates the following probable values for Hale Pohaku on the south side and Lake Waiau near the summit.

Consistent with the deduced values given above, are the following observed temperatures in ° F: (1) a maximum of 57.1 and a minimum of 18.9 during early August, 1935, at Lake Waiau; (2) a daily mean of about 50 for the latter half of July, 1939, at Hale Pohaku; (3) a noon temperature of 70 in the shade at 10,200 feet on July 1, 1939.

From the figures given, it seems probable that temperatures both above and below freezing occur practically every day in the year on the summit plateau. The results are seen in widespread development of stone stripes, *Polygonboden*, and other evidences of frost heave in the soil.

GLACIATION

GENERAL CONDITIONS

Proofs of glaciation.—Evidences of glaciation widespread on Mauna Kea above 11,000 feet are striated and plucked ledges, rounded roches moutonnées, moraines, and erratics (Pl. 1, figs. 1, 2; Pl. 2, fig. 2). Glacial

TABLE 2.—*Probable temperatures (in °F.) for Hale Pohaku and Lake Waiau*

Station	Hale Pohaku (9300 feet)	Lake Waiau (13,007 feet)
Mean annual	44	33
Mean January	40	29
Mean July	48	40
Lowest	14 to 18	0 to 10
Highest	70 to 80	.....

abrasion was generally rather feeble. In a few places the striated surfaces are sharply graved or even polished, but in the main they are only rudely grooved without conspicuous smoothing. Large faceted blocks on the more recent moraines generally show only faint striations. Nevertheless, over wide areas the results of glaciation are evident in the irregular terrane strewn with angular, somewhat faceted black, gray, and tan blocks of various sizes. Although these are all basic lavas with limited petrographic variations, they show nearly as many differences in color and general appearance as the truly diversified stones of a moraine in, for instance, the Southern Rocky Mountains.

Distribution, thickness, and movement of the ice.—During the latest glacial stage, here designated the Makanaka, ice covered approximately 26 square miles of the summit area and its margin lay generally at 10,750 to 12,000 feet (Fig. 3). Where retarded by cinder cones, the margin remained above 12,000 feet, but where valleys aided concentration of flow, the Makanaka glacier descended lower and in Pohakuloa Gulch reached about 10,200 feet. In the southwestern sector the ice margin was generally about 1000 feet lower (10,750 feet) than in the northern sector of the mountain. Conclusions regarding ice extent and movement published by Gregory and Wentworth in 1937 were generally confirmed during 1939, when the ice margin was first carefully mapped in detail.

The thickness of the ice in the summit area indicated by the height to which glacial blocks strew the sides of cinder cones which protruded as nunataks (Fig. 3). They indicate a maximum thickness during the Makanaka stage of about 350 feet, with an average of no more than 150

feet. The relatively feeble erosive effects in themselves suggest that the icecap was thin.

During the earlier stages the ice moved considerably beyond the Makanaka limit in some places but not in others. Present distribution of drift deposits indicates that the earlier icecaps were characterized by a marked extension of certain tongues and lobes, as in the vicinity of Pohakuloa Gulch on the southwest side of the mountain (Fig. 3). Due to concealment by later lavas, drifts, and forest cover, it was not feasible to map pre-Makanaka drifts areally during the time available in 1939. Figure 3 indicates exposures but not areal boundaries.

Steepness of mountain flanks.—The form of the mountain affected glacial movement. On the summit plateau gentle slopes favored accumulation of snow and its transformation into ice, but below 11,000 feet the generally steeper slopes must have greatly facilitated spreading and dissipation of such ice from upper margins. A strong development of marginal moraines near the 11,000-foot level during the latest stage and the distribution of earlier drifts at lower levels suggest that the icecap maintained a compact margin at or near 11,000 feet with projecting tongues or lobes that moved to lower levels in favorable sectors. Earlier glacial stages were characterized more by the vigorous extension of these few lobes than by expansion of the summit icecap as a whole.

Absence of cirques and confining valleys.—A notable feature of Mauna Kea glaciation is the absence of cirques and valley profiles characteristic of mountain glaciation. Lack of cirques reflects feebleness both of glacial action and of regional dissection favorable to the production of such features. The permeable character of the cinders and lava flows has led to a very high infiltration of precipitation and doubtless of glacial meltwater—as a result, fluvial erosion and valley deepening have been very slow. Moreover, lava flows have always tended to fill the radial valleys and thus heal erosional scars. This tendency of flows to follow and fill radial valleys on composite volcanic cones like Mauna Kea is so universal that few ravines show identical stratigraphic sections in both walls.

In a few places short stretches of upper valleys have been glacially smoothed. Rounded bottom profiles with steep walls occur but are probably not primarily due to glaciation. Postglacial erosion is generally inconspicuous. A U-shaped and glacially smoothed portion of upper Pohakuloa Gulch contains a narrow postglacial trench or slot some 10 feet deep and 20 feet wide. In upper Waikahalulu Gulch is a prominent breach some hundreds of feet wide in the Makanaka moraine (Pl. 1, fig. 3), but this was doubtless due largely to meltwater from the shrinking ice-

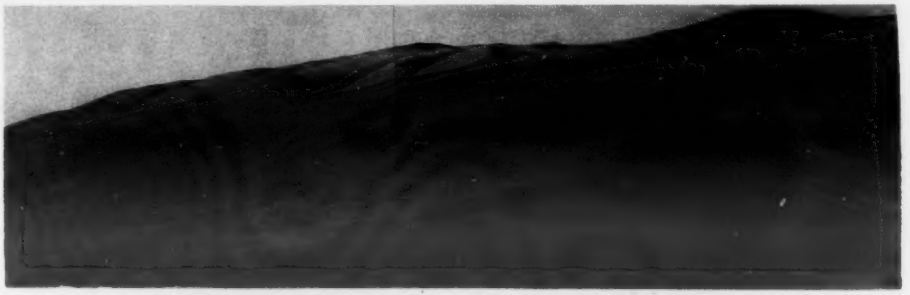

FIGURE 1. SOUTHWEST SLOPE OF MAUNA KEA VIEWED FROM PUU KALEPEAMOA
Center horizon approximates margin of summit plateau. Hale Pohaku is beyond right margin in scattered trees. Large dark cone on right horizon is Puu Keonehehee; below it, aa flows (dark) amid cinders; to left along horizon, light-colored Makanaka moraine. In shadow at left of view is west wall of Waikahalulu Gulch.

FIGURE 2. STRIATED AND FLUTED LAVA LEDGE
East of Waikahalulu channel at 12,000 feet.

FIGURE 3. VIEW TOWARD HEAD OF WAIKAHALULU GULCH FROM ABOUT 11,200 FEET
Three dark post-Makanaka flows have spilled into valley from east. Nearest cinder cone dates from same eruption as flows. Light bouldery areas are Makanaka drift.

MAUNA KEA SLOPE AND WAIKAHALULU GULCH

FIGURE 1. WEST WALL OF WAIKAHALULU GULCH
Waihu drift at top, overlying lavas that rest on Pohakuloa drift. Latter partly covered by talus but is seen to form two thirds of wall at left, where it rests on still older lava.

FIGURE 2. VIEW TOWARD PUU KEONEHEHEE (RIGHT) FROM EDGE WAIKAHALULU GULCH AT 9,700 FEET
Tongue of light-colored Makanaka drift in center distance partly covers broken aa flow at left and in foreground.

WAIKAHALULU GULCH AND PUU KEONEHEHEE

cap. The general feebleness of valley deepening by either ice or water indicates that the steep mountain slopes favored a type of sheet movement both of ice and meltwater, rather than concentration of flow in certain linear areas.

Discrimination between moraine and outwash.—The bouldery character of all drift deposits has in places rendered difficult the discrimination between ice-laid moraine and water-laid outwash. The Makanaka or last glacial stage left a heavy morainic belt composed largely of fresh, rudely rounded or faceted, sometimes striated boulders and stones, varying in size up to several yards in length. This moraine appears as a massive ridge, considerably steeper on its outer slope, sometimes attaining 25 degrees, than the mountainside at its foot. The crest, however, rarely is found to rise above the mountain slope leading upward toward the summit but rather appears as a mere flattening of the general 15- or even 20-degree declivity. Recognition of this moraine cannot confidently be made on the map alone, but field determination is easy. Small lateral moraines project from the Makanaka terminal moraine for short distances down Pohakuloa and Hanipoe gulches.

All other extramorainal accumulations of boulders are interpreted as outwash deposits from the Makanaka glacier or as older drift. General dissection or concealment of the original topographic forms of the older drift deposits lays heavy emphasis on their character as sediments. Whether a given deposit of boulders was laid in its present position by ice or by water is a decision that must sometimes be determined from several lines of evidence. Where the boulders are fairly uniformly of 1- or 2-foot sizes, or where there is evidence of rude stratification or arrangement in lenses, water was probably the transporting agent. If the boulders show a great range in size and if many are faceted and striated, if a fine matrix is present, if a considerable thickness—say 50 feet or more—is devoid of stratification, and if the deposit is associated with what appears to be a striated glacier pavement, the deposit may with some confidence be ascribed to ice. While many boulder beds seem to have been brought to the lower slopes of Mauna Kea by torrential flow of meltwater, the accumulated evidence leaves little doubt but that glaciers actually descended nearly to the south base of Mauna Kea at 6600 feet in the vicinity of Pohakuloa Gulch.

In many places glacial striae are unmistakable, occurring with long parallel flutings, chatter marks, plucked lee- and rounded stoss-edges. However, certain other smoothed or slightly scratched surfaces, found in or near stream channels, are due to stream scour. As can be seen in the more actively rounded parts of modern channels, the striae are short,

sporadic scratches, or scratches arranged in a spraylike pattern downstream from a small pit or obstruction on the surface. The general rock surface has a tortuous form, modified by overdeepened grooves and potholes, rather than the long parallel scorings of the typical glaciated surface. Small, obscure and weathered surfaces of this sort are difficult to discriminate from glaciated ones. Both types are known to occur on Mauna Kea.

SUCCESSION OF FOUR DRIFTS

General statement.—Evidences were found for four glaciations, each marked by heavy beds of till or tillite and separated from the next by lava flows sufficiently thick to indicate a considerable lapse of time. These glaciations are interpreted as stages, and have been named as follows:

4. Makanaka stage ..... latest
3. Waihu stage
2. Pohakuloa stage
1. Pre-Pohakuloa stage

The drift of the last stage is best exposed and most widely distributed. Some characteristics of the earlier drifts are obscure and can be interpreted only by analogy with the more fully revealed later drift. Age relationships are confusing where drift of one stage contains erratics derived from an earlier stage or is strewn with boulders by later outwash.

Makanaka drift and stage.—Drift of the last stage, the Makanaka, is named after the cinder cone Puu Makanaka (Fig. 3) against which is a conspicuous blocky moraine. The Makanaka moraine is also strongly developed for a mile or more on either side of Pohakuloa Gulch at 11,000 feet; it is crossed by the Humuula-Summit trail at about 11,300 feet one-third mile due west of Puu Keonehehee (Pl. 1, fig. 1) and is found on either side of Hanipoe Gulch at about 11,500 feet (Fig. 3). At most other places the margin of this drift is marked by a low ridge of fresh blocks that stand out sharply from the more stained and weathered lavas and older drift beyond. General features of the Makanaka drift have already been described in some detail by Gregory and Wentworth (1937). The fresh, comparatively unweathered, light-gray to tan rock fragments of this drift are outstanding. Most blocks are of fairly coarse-grained andesitic lava found extensively in the summit area. A less abundant rock type is the finer-grained black adze rock of Keanakakoi (The-cave-of-the-adze) at 12,500 feet, used so extensively by the native Hawaiians in adze manufacture. Over much of the area of both Makanaka and earlier drift are strewn spalls from adze-chipping operations carried on by natives who found the dense black adze rock in blocks transported

FIGURE 1. CHANNEL OF NAU GULCH ON NORTH OF MAUNA KEA AT 10,300 FEET
Shows pothole and other effects of stream scour. Illustrates shallow depth of most water courses at this level.

FIGURE 2. SNOWBANK FOUR FEET DEEP ON INNER SLOPE OF SOUTH RIM ON SUMMIT CONE
Oriented sun pitting. July 24, 1939.

NAU GULCH AND SNOWBANK ON MAUNA KEA

FIGURE 1. TYPICAL WAIHU DRIFT
Brownish-crusted, black exfoliated boulders, at 9900 feet northwest of Waihu Springs.

FIGURE 2. WAIHU (?) BOULDERY DRIFT BENEATH LAVA IN EAST WALL WAIKAHALULU GULCH AT 10,200 FEET
Boulders soled and some faintly striated.

WAIHU DRIFT

downslope. With the spalls are found distinctive vesicular, ball-shaped hammer stones.

Little true till of Makanaka age is exposed. Much of the surface has been so washed by water that only a coarse lag of blocks and slabs is now present. Several seepage springs in the morainic slopes near Pohakuloa Gulch are due to less pervious drift below the bouldery surface zone, and rather clayey gray till is exposed in the walls of Pohakuloa Gulch at the 11,000-foot level.

Waihu drift and stage.—At many places outside of and below the limiting light-colored Makanaka moraine, there are accumulations of boulders unrelated to adjacent flows and aa debris but evidently erratic. Everywhere the colors are browns, deep grays, and purples, darker than the Makanaka drift. In several places, as on the west side of Waikahalulu Gulch between 10,000 and 10,250 feet, the boulders are cemented into a brown tuffaceous matrix. This drift, clearly older than the Makanaka, has been named the Waihu from prominent exposures at about 9700 feet near Waihu Springs west of Pohakuloa Gulch (Pl. 4, fig. 1).

The distinct character of this drift is also indicated by its color difference as seen from the southern foot of Mauna Kea, near Pohakuloa Camp, for instance. In late afternoon when the light gray and tan Makanaka moraine is conspicuous on the high shoulder of the mountain, a ragged fringe of darker color but similar texture is visible below. Traverses along the mountain slope below the Makanaka moraine demonstrate that this darker brown fringe is a distinct and older drift, in places strewn down the mountainside to elevations as low as 8000 feet; it is not to be confused with the still darker aa flows and cinders.

Much of the Waihu drift contains distinctive dark-gray to black, fine-grained basalt derived from aa flows widely distributed over the mountain below the Makanaka moraine. This basalt is peculiarly susceptible to spalling. Many of the Waihu boulders are chunky, somewhat rounded masses, derived from contraction blocks of the lava flow. Their glacial surface is tan to dark brown with a somewhat etched texture, but commonly this older surface is partly removed by exfoliation showing the black unweathered interior of the rock and strewing the ground with curved spalls. Thus the blocks have become progressively rounded. Lava flows of the same basalt are less markedly affected by this exfoliation than glacial blocks, probably because of their very irregular and somewhat honeycombed flow surface.

Pohakuloa and pre-Pohakuloa drifts and stages.—Both stratigraphic and areal relationships indicate strongly the existence of two distinct drifts representing glacial stages prior to the Waihu. Best exposures of these

earlier drifts occur in the Pohakuloa and Waikahalulu gulches and the intervening area. The nature of the evidence is shown in Figure 4, a diagram in which conventionalized contours are employed to give depth to the gulches and hence to permit the presentation of vertical as well as

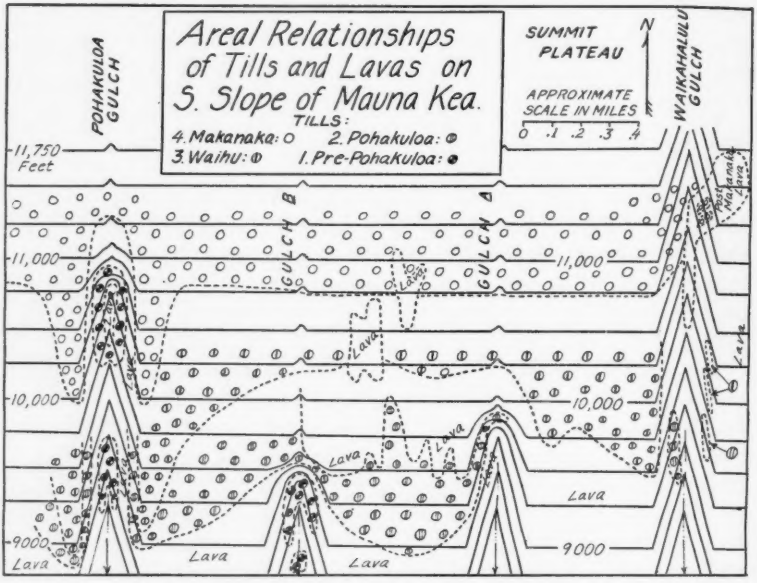

FIGURE 4.—*Idealized areal map on contour base*
Showing relationships of tills and lavas on south side of Mauna Kea.

horizontal relationships. Field study is complicated by (1) the concealment of the drifts by soil and forest cover and lavas, save in the deeper gulches, and (2) the characteristic tread-and-riser type of stream profile which leads to small isolated outcrops in the deeper parts of stream valleys. Till bodies when once pierced at some point in a stream bed are readily eroded back to form dry waterfalls which alternate with channel stretches of relatively gentle gradient. The complete areal pattern of the several drifts was not worked out in the present survey, nor could it be without much detailed study. If the reader has the Mauna Kea and Waikii topographic quadrangles at hand, the evidence here presented may be followed to better advantage than with Figure 4 alone.

Areal and stratigraphic relationships (Fig. 4) indicate two drifts older than the Waihu. On the south slope the Makanaka moraine extends

down to approximately the 10,750-foot level, save along Pohakuloa Gulch where it projects in two narrow lateral moraines down to about 10,000 feet. The Waihu moraine sends down laterals to nearly 9000 feet along the sides of Pohakuloa Gulch but farther east descends only to the vicinity of the 10,000-foot contour. Here it is rather complexly related to various flow units of dense black basalt, some of which occurs as boulders in the Waihu drift. Thus these black basalt flows in part, at least, antedate the Waihu glacial stage. Still farther east a heavy Waihu moraine occurs on the west side of Waikahalulu Gulch at 10,000 feet (Pl. 2, fig. 1). This drift contains numerous angular, well-faceted and striated boulders firmly indurated in a brownish-red tuffaceous matrix. The maximum thickness is about 75 feet. This drift thins out toward the south but extends down to approximately 9500 feet. Where the gulch rim is about 9700 feet in altitude, the west wall of Waikahalulu Gulch shows this section:

	Thickness (in feet)
Waihu till	30
Lavas, about 10 flows totaling	35–40
Boulder bed, rudely layered; contains both faceted and somewhat rounded boulders up to 5 feet in size, set in gray tuffaceous matrix, the whole standing in nearly vertical wall. A till	90
Talus	60
Total	215–220

Both upstream and downstream the lower till bed lenses out, and the lavas above and below it thicken. This till is clearly older than the Waihu, and is tentatively assigned to the second glacial stage herein named the Pohakuloa. On the east side of the gulch is a thinner mass of very similar indurated drift at the same level, probably identical in age. Higher up on the east gulch wall, where the rim is about 10,200 feet, three boulder beds are interbedded with lavas (Pl. 4, fig. 2). Each boulder bed is only a few feet thick and consists of subangular and striated boulders set in a fine tuff matrix. The upper 3 to 4 feet of each boulder bed is stained pinkish to brick-red, presumably by the heat of the flows above. These beds are till of somewhat uncertain age but are here correlated with the near-by Waihu drift to the west.

At the 9750-foot level in a gulch .8 of a mile west of Waikahalulu is a dry waterfall with a rim composed of dense basalt flows of the pre-Waihu type. Below 25 feet of these lavas is a 5-foot bed of boulders, faceted and striated, set in an ash matrix and evidently a till (Fig. 4, Gulch A). Followed southward and westward this till thickens and emerges from its basalt cover to form the surface deposit of a considerable area. At one point 100 feet is exposed. Boulders up to 5 feet in size are set in a fine

mauve-gray matrix so firmly indurated that pebbles are loosened only by vigorous pounding with a pick. This is Pohakuloa drift. Farther westward but .3 of a mile east of Pohakuloa Gulch this till rests on from 20 to 40 feet of lavas, below which is a still older 70-foot boulder bed (Fig. 4, Gulch B). This deposit is the earliest or pre-Pohakuloa drift. Although showing some evidence of stratification at certain horizons, it is generally similar to the other pre-Makanaka tills or tillites. Boulders up to 4 feet in size, many of them faceted and some striated, are rather well indurated in a gray tuffaceous matrix.

Westward from Gulch B (Fig. 4) the pre-Pohakuloa drift is covered by lavas, but the Pohakuloa forms the surface deposit until overlapped by the large bouldery lateral moraine of Waihu age near Pohakuloa Gulch. Along this gulch, between rim elevations of 8900 to 10,500 feet, the upper part of the gulch wall consists of either Waihu or Pohakuloa drift or both, resting on lavas below which is a bed of pre-Pohakuloa till. This till body is not continuously exposed, but appears at several points under dry waterfalls, where the stair-step stream profile has penetrated most deeply into the sloping strata of the volcano. The boulders and blocks in this pre-Pohakuloa till (or rather, tillite) are largely of rather vesicular basalt from the underlying flows. The vesicular texture and olivine phenocrysts of these fragments effectively distinguish this drift from the three lying stratigraphically above it. A typical section occurs where the rim of Pohakuloa Gulch has an elevation of 9620 feet:

	Thickness (in feet)
Nonindurated, bouldery moraine, Waihu drift	about 25
Indurated mauve to gray bouldery till, Pohakuloa (?)	about 50
Lava flows, 16 in all	90
Boulder bed, large boulders, many faceted, set firmly in gray tuffaceous matrix, pre-Pohakuloa	90
Total	255

On other parts of Mauna Kea where Pleistocene stratigraphy is less complete, age relationships of the older drifts are more difficult to work out, and convincing evidence is generally lacking. A heavy bouldery till resembling the Waihu occurs beneath lavas down to an elevation of 7600 feet west of Pohakuloa Gulch, and a gray till resembling pre-Pohakuloa occurs as low as 6900 feet near Pohakuloa Gulch (Fig. 3). The fact that pre-Pohakuloa till is usually the thicker of the two, when it occurs with Pohakuloa at higher levels, indicates that the pre-Pohakuloa glaciation was the more vigorous. For this reason these low-level deposits are tentatively assigned to the pre-Pohakuloa stage.

CORRELATION

Correlation with glacial stages elsewhere.—The writers believe that the four drifts found in Hawaii correlate with the four major glacial stages recognized in North America and Europe, as follows:

Stage No.	Hawaii	North America	Alps
4	Makanaka	Wisconsin	Würm
3	Waihu	Illinoian	Riss
2	Pohakuloa	Kansan	Mindel
1	Pre-Pohakuloa	Nebraskan	Günz

Although this correlation rests mainly on correspondence in number of stages, there is strong evidence that several stages are represented by the drifts in Hawaii. The two oldest drifts are firmly indurated and deeply oxidized, as are drifts of the older Pleistocene elsewhere. Between the Hawaiian drifts are thick lava series which record intermittent outflows of considerable duration. These facts are consistent with the interpretation of the several drifts as separate stages rather than interstadial deposits. Several types of evidence usually associated with multiple glaciation are conspicuously absent. Soil profiles on the older drifts were not recognized. Their absence is readily explained by the relatively slow, shallow weathering characteristic of these altitudes in Hawaii, plus the anamorphic effect of the superincumbent lavas which have developed reddish, baked zones several feet deep. No fossil evidence was found associated with the older drifts. No evidence was discovered pointing toward warmer interglacial climatic conditions, though these must have existed. There are no indications of deep dissection of Mauna Kea during interglacial stages—rather, the thin yet numerous flows between glacial stages sought the valleys and gulches and thus healed these breaks in the great lava dome. Despite the absence of compelling proof, the facts warrant a tentative correlation of the four drifts on Mauna Kea with the four glacial stages generally recognized elsewhere.

Possible correlation with Quaternary marine terraces.—The possible correlation of glacial and interglacial stages on Mauna Kea with Pleistocene strand-line levels as worked out by Stearns for the Islands of Oahu, Maui, and Lanai (Stearns, 1935; 1938), Daly (1934), and others, at once suggests itself. Stearns recognized 11 positions of sea level from +250 to —300 feet that in part correspond closely with those of Cooke on the Atlantic coast of the United States, as well as others at —1200 feet and +1200 feet (Stearns, 1935, p. 1952, 1930-1931; 1938, p. 618-620). The existence of these former shifts in sea level is adequately supported, and the first group of 11 are very probably due to glacial control during the

Pleistocene. Unfortunately, the possibility of direct correlation on the Island of Hawaii between glacial stages and negative sea-level stands seems remote. Only the mountain masses of Mauna Kea and Kohala are old enough to carry such records. Much of the Hamakua (northeast) coast has been so actively eroded in recent time that any evidence has probably been destroyed. Critical study of the deep Waipio canyons in the north may reveal an erosional history partly correlative with the glacial history. Lack of graded or erosionally unified valleys extending from summit area to sea level and the slight erosion of the entire dome greatly reduce the possibility of locally deciphering the relationship between glacial history and shoreline history, even if and when some tangible shoreline sequence is worked out on the more ancient parts of the island of Hawaii.

RELATION OF GLACIATION TO VOLCANIC HISTORY

MAKANAKA STAGE AND VOLCANIC ACTIVITY

Most cinder cones of the summit area appear to be older than the latest glacial deposits. Certain cones are strewn with large, light-colored blocks, a fact leading to speculation as to whether the blocks are fragments from an older drift incorporated by the eruption into the structure of the cone. The alternative interpretation is that after the eruption these blocks have been placed by glacial ice which moved against or over the cones. Evidence generally favors the latter interpretation. In every instance the blocks are more numerous and at a higher level on the side against which the ice moved, or are distributed over both top and sides of low cones which thus appear to have been surmounted by the glacier. Small glacial debris is rarely associated with large glacial blocks on the side slopes, a situation unquestionably due to the universal creep, perhaps largely under winter snows. By this process fine debris is carried down with cinders, leaving the larger, more deeply sunken blocks nearer their original positions. For the same reason bombs on cinder cones also lag behind while finer cinders move downward.

POSTGLACIAL ERUPTIONS

Very little eruptive activity has occurred on Mauna Kea since the Makanaka glacial stage. The most conspicuous exceptions are two small unnamed cinder cones near the head of Waikahalulu Gulch northwest of Puu Keonehehee, associated with a small valley flow of aa that extends approximately 1 mile down the gulch (Figs. 3, 4; Pl. 1, fig. 3). At this point Waikahalulu Gulch forms a deep breach in the heavy Makanaka moraine, probably cut largely by meltwater from the shrinking icecap. Just within the moraine two cinder cones have erupted and from them

aa flows have spilled down into the gulch from the east at three points. At many places within the gulch the overlapping relationship of this aa flow to the Makanaka drift is shown. A small cone and flow upslope from Puu Makanaka bear similar though somewhat less obvious relations to the Makanaka drift. At various places on the downslope side of the Makanaka moraine there are tongues of aa some of which are almost certainly post-Makanaka. However, due to the steep slopes and the loose, detrital character of both aa and boulder moraine, much shifting has resulted from creep, and positive proof of stratigraphic relationships is difficult to establish.

GROWTH OF MAUNA KEA DURING THE PLEISTOCENE

A consideration of the volumetric relationships of the lavas associated with the several drifts on Mauna Kea leads to the conclusion that most of this mountain's growth occurred prior to the ice age. A comparison with Mauna Loa is instructive. The latter's fresh, almost uneroded form and immaturity of soils and forest reflect its steady upbuilding by lava outpourings up to the present time. During the past 100 years Mauna Loa has extruded lava at a rate estimated between 5000 and 10,000 millions of cubic meters per century; this amounts to an average growth over the entire area of the dome of 3 to 6 feet per century. If this rate has been sustained for the entire postglacial period of, let us say, 30,000 years, Mauna Loa must have been 1000 to 2000 feet lower than now at the end of the ice age and was presumably lower still during that period. However, on Mauna Kea the almost negligible post-Makanaka flows probably average no more than 10 feet thick over their estimated area of 1 square mile, or an average increment of 0.01 foot to the whole surface of the dome, roughly 1000 square miles. If post-Makanaka time be estimated at 30,000 years, the average rate of upbuilding applied to the whole dome is approximately 0.000,033 foot per century. Post-Waihu—pre-Makanaka flows are estimated to cover 5 square miles to an average depth of 10 feet, or an average increment to the entire Mauna Kea dome of 0.05 foot. Flows of the interglacial stage preceding the Waihu are estimated at 10 square miles and 30 feet in thickness, giving an average increment to the entire dome of 0.3 foot. Flows overlying the oldest drift (pre-Pohakuloa) are estimated to cover 25 square miles with an average depth of 100 feet, giving an average increment to the entire dome of 2.5 feet. To the total of these flows must be added the volume of the summit cinder cones estimated at 0.5 cubic mile (Wentworth, 1938), and the widely dispersed Waiau ash formation estimated at 3 to 10 cubic miles. The total volume of these Pleistocene lavas, if we take the larger figure for the Waiau ash, is slightly more than 11 cubic miles. This is an average

increment to the entire dome of approximately 58 feet, or 0.0058 foot per century if the Pleistocene epoch be regarded as 1,000,000 years. Although the figures used herein are chiefly estimates and might be modified one to four times in either direction by other computers, they nevertheless indicate orders of magnitude. By any approach the growth rate of Mauna Kea during the Pleistocene appears very small, and indicates that the mountain had almost attained its present dimensions before the glaciers appeared.

On the other hand, the inferior height possessed by Mauna Loa at the end of the ice age renders it doubtful that this mountain could have supported an icecap. If it did, the latest glacial deposits are now presumably buried under 1000 to 2000 feet of lava.

POSTGLACIAL WEATHERING AND EROSION

It is unfortunate that data as to rates of soil formation cannot be developed from present observations. In many continental areas, interglacial periods were characterized by climate conditions very different from those during glaciation, and distinctive soils were developed on the previously deposited glacial drifts. Had Mauna Kea glaciation been vigorous enough to descend to levels of 2000 or 3000 feet above sea level where climate is humid, there would have been abundant soil formation on the drifts, which generally lack rocks resistant to intense chemical weathering.

However, with glacial till only at elevations above 7000 and chiefly above 10,000 feet, weathering even of the earlier drifts is very slight, and is confined chiefly to surface decomposition, oxidation, and hydration of the outer shells of the boulders. From sea level to the summit of Mauna Kea an extraordinary variation in extent of weathering exists. Up to heights of 1000 to 2000 feet where rainfall is 100 to 250 inches annually, weathering has produced red-brown lateritic soils. At greater elevations the soils are progressively less well developed and at 6000 to 9000 feet consist largely of volcanic sand and cinders modified only by the addition of small amounts of organic material. From 9000 feet to the summit the surface materials are cinders and fragments of slightly weathered basalt affected mainly by freezing and thawing (Foster, 1939). Where drift remnants now lie, the weathering conditions are generally those of cool arid countries, with little tendency toward the chemical processes of lateritization. The climate of the Makanaka drift area is periglacial today, with temperatures below freezing at night during most of the year. No significant soil cover or forest growth has ever developed here or should be expected to develop under existing conditions.

Present condition of the drifts can be described only in general terms. The Makanaka boulders and finer materials are light-gray and cream-

colored. Freshly broken boulders show a very slight bleaching of the surface to a depth of 1 or 2 millimeters, probably a slight kaolinization which contrasts with the darker gray of the fresh rock. Signs of oxidation are rare, and in many places detritus due to frost action shows no evidence of alteration.

The Waihu drift, where it has been most generally recognized, consists largely of the fragments of a certain series of post-Pohakuloa flows peculiarly susceptible to exfoliation. The boulders of the drift are medium brown to buff or cream on the outside, but this somewhat weathered and oxidized surface is actively scaling, due in part and perhaps largely to extreme temperature changes. Several times at the 10,000-foot level the writers experienced squalls of driving, chilly rain or sleet, after bright sunny midday periods when rock surfaces were heated to 140° or even 160° F. In some places the tuffaceous drift matrix is oxidized buff or brown, but in others it is light gray with no sign of oxidation. The surfaces of aa flows and fragments as old as or older than the Waihu drift are commonly brown or red to depths of only a few millimeters. These, however, may have been oxidized by residual heat or under atmospheric conditions related to the eruption and thus do not indicate rate of surface oxidation under ordinary conditions.

The Pohakuloa drift occurs in two forms. Thin layers which outcrop at many points under eroded edges of post-Pohakuloa flows usually are deep pink or red. Both cobbles and matrix are thus colored. The relations and fading downward indicate that this was produced by the heat of the overlying flow. Farther away from the lava the colors are gray, drab, or purple. These same colors also prevail in the thick beds of Pohakuloa tillite.

In many such outcrops a certain amount of color change indicating oxidation is accompanied by firm induration, suggesting the same combination of katamorphic and anamorphic effects seen in the older alluvium formations of drier parts of Oahu. These alterations collectively indicate that the Pohakuloa tillite is definitely older than Waihu drift, but Hawaiian weathering in its relation to elevation is not sufficiently well known to permit a useful discussion.

Little is known of weathering effects on the pre-Pohakuloa tillite. Most exposures are of a horizon well below the top of the drift body and have been laid open to atmospheric effects only in the present stage of erosion. No marked weathered zone has been observed at the top of this drift.

Lack of extensive weathering, of thick soil beds, or of evidences of plant growth between the several epochs is due to no single cause. While temperature conditions at these levels have probably never favored chemical

weathering, it is also likely that prevailing aridity, very rapid soil drainage, steep slopes, and lack of vegetation have all conspired to prevent development of significant soil covers, even over very long periods. Thus chronological interpretations by evidence of this kind have generally been defeated.

SUMMARY AND CONCLUSIONS

Four glacial stages on Mauna Kea are indicated by four distinct drifts separated by series of lava flows up to 100 feet in total thickness. Such criteria as progressively greater internal induration, greater surface weathering where broadly exposed, and development of soil and plant cover on the older drifts indicate that true glacial stages are represented. A tentative correlation is indicated with the Nebraskan-Günz, Kansan-Mindel, Illinoian-Riss, and Wisconsin-Würm stages of North America and Europe. No other known island summit in the Pacific area affords evidence of this sort. It is highly significant that glaciation occurred here on a summit where no permanent snow fields exist today. If Mauna Kea were higher than its present 13,784 feet, the snow line today would probably lie between 14,000 and 15,000 feet. During the Pleistocene, general ocean-wide climatic changes led to periodic fall of the snow line below the level of 13,000 feet, the probable approximate height of Mauna Kea early in the Pleistocene. Glacial advances in some continental areas may be correctly attributed to increases in precipitation, but in Hawaii this explanation seems less plausible than ocean-wide lowering of temperature. Most of the cinder cones on Mauna Kea probably were erupted during the Pleistocene, but their relations to moraines indicate that all but three or four were formed prior to the last glacial stage. Upbuilding of the volcanic dome by lava outflows has been negligible since glaciation, and even since the beginning of the ice age it has not exceeded 100 feet for the whole dome.

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MAMMALIAN FAUNA FROM THE PLIOCENE OF HONDURAS

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ABSTRACT

A small but interesting mammalian fauna has been obtained from the Mejocote Valley in the vicinity of Gracias, Honduras, from a series of beds of sand, clay, and volcanic ash, a formation to which the name Gracias is given. The fauna, which consists of *Amphicyon*, *Pliohippus*, *Neohipparion*, *Procamelus*, *Blickotherium*, and an indeterminate cervid, appears to be lower Pliocene in age. Horses are the most important element, and it is mainly on the basis of these forms that the age deter-

mination has been made. A small flora, associated with the mammalian fauna, suggests that the animals lived in or on the edge of a subtropical rain forest. The absence of mammals of South American origin strongly suggests that no land connection existed between the two Americas during the lower Pliocene, and the presence of the northern type fauna indicates that no marine portal was present at the isthmus of Tehuantepec during the time of deposition of the Gracias formation.

INTRODUCTION

During the winter of 1937–1938, a party from Walker Museum of The University of Chicago undertook an expedition to the Department of Gracias, Honduras, for the purpose of investigating certain localities from which fossil mammals had been reported. This trip was made possible by a grant from the Penrose Bequest of The Geological Society of America and by support from The University of Chicago. The personnel consisted of Paul C. Miller, Paul O. McGrew, and Grayson E. Meade. About 2 months were spent in actual field investigation.

Fossil mammals have been reported from Honduras on three previous occasions. Leidy (1859) described a tooth of a mastodon from the vicinity of Tambla as *Mastodon ohioticus* and later (1869) published a figure of this tooth. Nason (1887) reported the occurrence of numerous bones of fossil mammals, presumably mastodon, in the Department of Yoro in the Nigrito Valley about 27 miles east of the Ulua River.[1] Owing to the difficulty of transportation, he was unable to carry these fossils out, and they were not described. Frick (1929) described and figured two mastodons, *Blickotherium blicki* and *Aybelodon hondurensis*, from the vicinity of Tapasuna [2] and listed a fauna based on fragmental material. From this locality he identified the following forms:

Hyaenognathus	*Protohippus*
Pliohippus	Teleocerine rhinoceros
Hipparion	Procameline
Merychippus	

Frick considered the formation in which these forms appear to be Upper Pliocene in age.

Two regions were visited by the party from Walker Museum—one near the town of Gracias and a second near the small village of Tapasuna. The locality near Gracias provided the better fauna because of larger exposures and more fossiliferous strata. One small area of

[1] The spelling of names and designation of rivers given by Nason is somewhat different from that found on standard maps at the present time. He referred to the Ulua River as the Ulna. The Coymayagua River, upon which the locality occurs, is called Cuyamapa on some maps. The course of this river is plotted somewhat differently on various maps. The fossil locality appears to lie about 15 miles above the junction of the Ulua and Coymayagua rivers.

[2] Spelled Tapasuma by Frick.

badlands in this area, the Rancho Lobo locality, yielded a good collection, but away from this spot only isolated horse teeth and fragmental mastodon remains were obtained. Because of the time required for collecting and the relatively short duration of the stay in the field, observations of a purely geological nature were necessarily limited to the immediate vicinity of regions in which fossils were to be expected.

The writers wish to express their gratitude to the Honduras governmental authorities for permitting them to carry out the work and for their co-operation in many other ways.

GENERAL GEOLOGY

Although numerous publications dealing with the general geology of the Republic of Honduras exist, there is little detailed information available concerning the physiography, stratigraphy, and geologic structure of the country. The foundations of our present knowledge were laid by Sappar (1899; 1905a; 1905b) as a result of his extensive travels through the Republic and adjacent countries. His work has been supplemented and re-interpreted by a number of subsequent writers. Powers (1918) reported on the geology of Guatemala and Honduras, dealing with the stratigraphy, structure, and tectonics of the region. Redfield (1923) published a short report on the petroleum possibilities of Honduras in which he summarized the data available at that time and added certain facts and interpretations. Bengston (1926) wrote a brief article on the physiography of Honduras in which he outlined the physiographic divisions and described each briefly. The work which had been done up to 1935 was summarized by Schuchert (1935). By a co-ordination of the available data he was able to offer a clearer picture of the geology of this part of Central America than had been possible up to that time and to develop a number of interesting concepts of the processes and steps which led to the conditions prevailing at the present time. There have been a number of other short papers dealing with various details, but most of the information on the geology of Central America, and particularly Honduras, can be obtained from the references cited.

The primary objectives of the expedition sent into Honduras from Walker Museum were the collecting of vertebrate fossils from Tertiary deposits in the vicinity of Gracias and the study of the formations in which these fossils occur. Time for careful study of many unsolved geological problems was not available. It is felt, however, that it is advisable to present, in this paper, a short summary of the general geology of Honduras as a preface to a more detailed account of the conditions which exist in the valley of the Mejocote River near Gracias

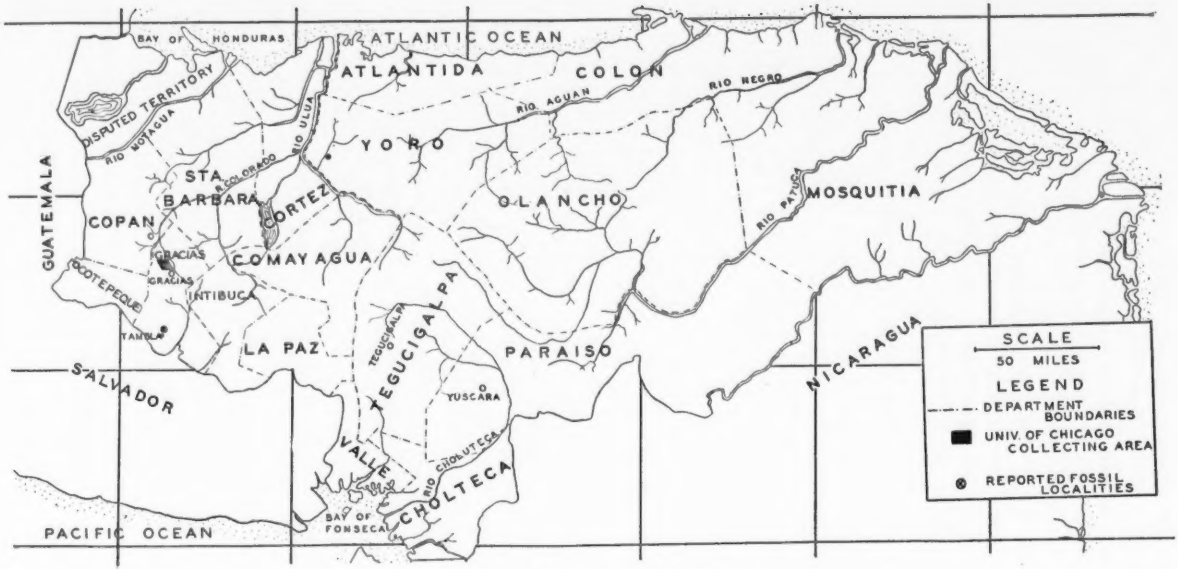


FIGURE 1.—*Outline map of Honduras*

Indicating area studied and previously reported fossil localities.

in which the present work was carried on. The information used in the summary was derived chiefly from the literature.

Honduras lies in the northern or nuclear portion of Central America, a region characterized by strongly folded and faulted mountains with a general east-west strike. Structurally it is related to Chiapas, Guatemala, British Honduras, and Northern Nicaragua. Much of the country is extremely rugged, and travel, except by airplane, is difficult. Many parts are heavily overgrown with vegetation, and geological work is rendered difficult by the lack of exposures.

Bengston (1926) recognized four distinct physiographic divisions of Honduras: (1) the coastal and alluvial plain confined largely to the Caribbean coast and the lower parts of the valleys of the rivers which empty into the Caribbean Sea; (2) the lower terrace plains developed along the river courses; (3) the higher terrace plains developed mainly in the upper reaches of the rivers; and (4) upland comprising the greater part of the country. It is in the last-named division that the fossiliferous deposits visited are located.

The drainage of Honduras is largely to the north and east into the Caribbean Sea. Along the southern margin of the country, extending somewhat farther north in the central and western parts, are streams which flow southward into Nicaragua, the Bay of Fonseca, and San Salvador. In the extreme west a few small streams flow westward into Guatemala.

Two principal mountain ranges have been recognized in Honduras. The southern range, called the Southern Cordilleras by Bengston, is composed of the Depilto, Yuscaran, San Juancito, Opatoro, and Opalaca Mountains. It is in general higher than the northern range, the Northern Cordilleras. For much of its length it forms the divide between the northern and southern drainage system. The Choluteca River, however, flows southward from north of Tegucigalpa and heads partially on the northern side of the range, flowing toward the Bay of Fonseca between the Depilto and Yuscaran Mountains. Between the two ranges are rugged highlands marked by transverse ranges and dissected by rivers. A striking physiographic feature of Honduras is the Great Honduras Depression which forms a line of communication between the Bay of Fonseca on the Pacific and Puerto Cortez on the Caribbean. This depression marks an interruption of the east-west structural trend of the mountains and has many structures of north-south strike. Redfield (1923) suggests that it may be a graben.

Information concerning the stratigraphy of Honduras comes largely from the work of Sappar. Later work has altered a few of his age de-

terminations but has not produced sufficient evidence to permit accurate dating of many formations. Gneisses, schists, phyllites, and slates appear to be the oldest rocks exposed in the country. They are pre-Mesozoic and generally have been referred to the pre-Cambrian. Definite data on their age are not available. Old granites are widespread in the mountain ranges, but their exact age has not yet been determined. Paleozoic sedimentary rocks appear to be poorly developed. Limestones, which underlie the plant-bearing beds at San Juancito, have been called Carboniferous by Newberry (1888), but evidence for this assignment appears to be inconclusive. The overlying beds, comprising the Tegucigalpa formation, have been identified as Rhaetic by Newberry (1888) on the basis of their cycad flora. Knowlton (1918) has confirmed this identification but states that there is some possibility that they are Jurassic. No beds thus far identified can be referred to the Jurassic with absolute certainty. Both lower and upper Cretaceous marine beds were identified by Sappar (1905). They are best developed in northwestern Honduras; no beds of this age have been identified in the eastern or southern parts of the country. Rocks which definitely can be assigned to the Eocene or Oligocene have not been found, although the "Old Cenozoic" of Sappar (1905) may belong to one or the other or to both. Marine Miocene beds are known in northwestern Honduras. Marine Pliocene formations occur along the Ulua River. Some clays, breccias, and conglomerates along the Rio Coco may represent fresh-water Pliocene, but, insofar as known, these beds are unfossiliferous. Much of the southern and western part of Honduras is covered by vast sheets of eruptive rock, which appear to be at least in part Pliocene. Fossiliferous continental deposits of lower Pliocene age occur in the vicinity of Gracias. It is from these that the fauna and flora described in this paper were taken. The beds appear to be quite extensive. Continental Pleistocene deposits have been reported from several localities, and it is probable that they are rather widespread. Marine Pleistocene occurs in northeastern Honduras.

Sappar (1905) believed that there were three periods of orogeny in Honduras: one during or at the close of the pre-Cambrian, a second at the close of the Paleozoic, and a third during the Miocene. Powers (1918) points out a fourth which occurred in the late Pliocene or Pleistocene. Evidence of this last orogeny appears in the Pliocene beds studied. The dips imposed upon these beds are quite varied and suggest a torsional movement. Vertical uplift is suggested by terracing which is clearly shown in the region studied. These features are discussed further in the succeeding section.

GEOLOGY OF MEJOCOTE VALLEY

GENERAL GEOLOGY

Extending northward for about 40 miles from the town of La Campa in the Department of Gracias is the valley of the Mejocote River. To the east and west of the valley rise high, mountainous ridges covered with a dense growth of vegetation. In the valley, close to the Mejocote River and its tributaries, is located a number of towns and villages, the largest of which is Gracias, the capitol of the Department of Gracias. The elevation of this town is about 2500 feet, but in its vicinity peaks rise to elevations of over 7000 feet. The composition of the high mountains is unknown, but from boulders and cobbles carried into the valley by streams, it is apparent that crystalline intrusives are extensively exposed.

The floor of the valley is very rough and is covered for the most part by a dense growth of vegetation. Exposures of underlying rock are few and their extent is limited. In all places where examination of the rock was possible, the deposits were found to be composed of clay, sandstone, volcanic ash, and conglomerate. In a few places thin beds of marl were present. The beds range in thickness from over 600 feet to 1000 feet. Erosion has been active in the valley since the deposition of these beds and has cut to successively lower levels in such a way as to produce a series of terraces, due in part, at least, to the presence of relatively resistant layers and perhaps to successive uplifts of the valley floor as well.

Only in very steep-sided cuts where small tributaries are eroding headward rapidly are there exposures of any considerable size. One of these cuts, that at Rancho Lobo, 3 miles north of the town of Gracias, has yielded a considerable portion of the vertebrate fossils described in this report. Most of the other exposures visited are merely small, bare patches on the sides of hills. The latter type of exposure is characteristic of the Tapasuna-Mangual locality about 10 miles north of Rancho Lobo. The dip and strike of these river-laid deposits vary greatly over short distances. For example, at Rancho Lobo the dip is 7° NE. and but a few miles farther north, in one small area, it is 30° S. Similar rapid variations were noted elsewhere.

The age of these alluvial deposits, on the basis of the contained vertebrate fauna, appears to be lower Pliocene. The horses were of the greatest value in arriving at this determination. The fauna upon which the age is based is as follows:

Pliohippus hondurensis n. sp.
Neohipparion montezuma (Leidy)
Procamelus sp.
Cervid indet.
Amphicyon sp.
Blickotherium blicki Frick

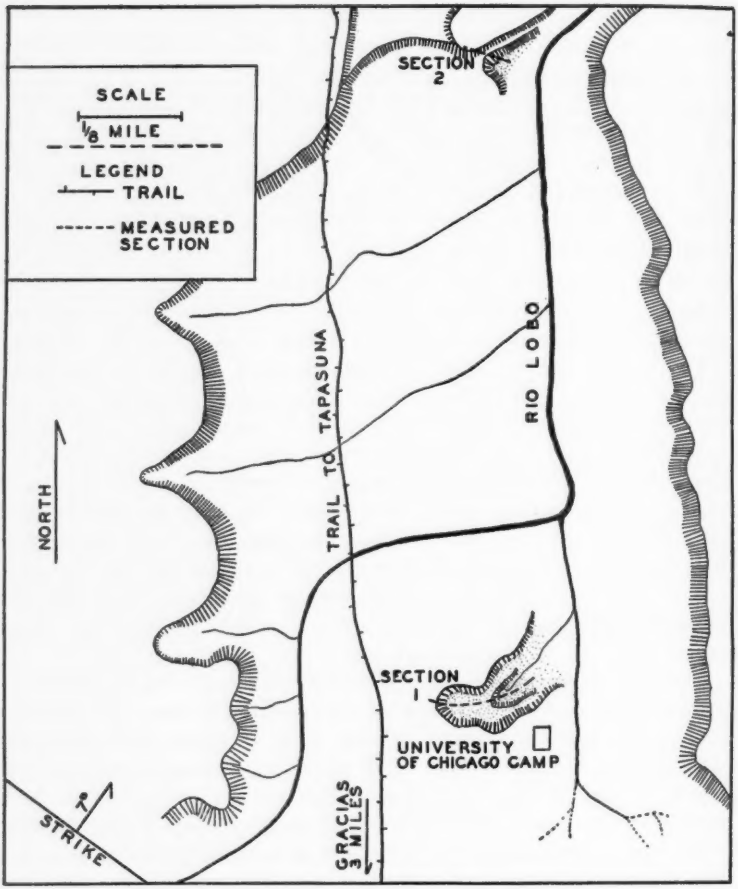


FIGURE 2.—*Outline map of Rancho Lobo locality*

Fossils of no other age were found in the area, although a vertical range of over 600 feet was covered during the exploration. In all exposures in which fossils were found, and fragments are not uncommon, the remains could be identified as pertaining to one or more of the genera found in the richer deposits of Rancho Lobo or Tapasuna. It appears from this that the exposed Tertiary beds in the vicinity of Gracias may be considered as comprising a single, lower Pliocene formation for which the name of Gracias formation is suggested. The type sec-

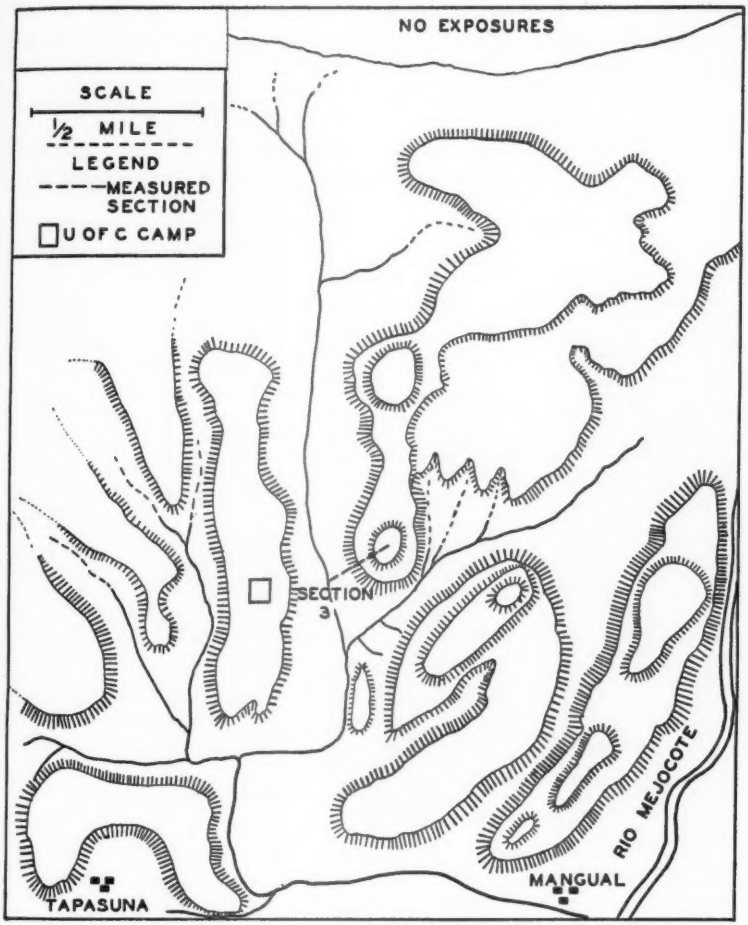


FIGURE 3.—*Outline map of Tapasuna-Mangual locality*

tion is designated as the exposure at Rancho Lobo, 3 miles north of Gracias (Section 1).

DETAILED GEOLOGY OF FOSSIL LOCALITIES

Rancho Lobo locality.—This locality (Fig. 2), located 3 miles north of Gracias on the trail to Tapasuna, was the most productive area visited. The best exposures occur in the banks of a small tributary of the Lobo

River just east of the trail. Rapid erosion is in progress, and the exposures more nearly approach badland conditions than do any of the others visited. The thickness and sequence of beds is shown in section

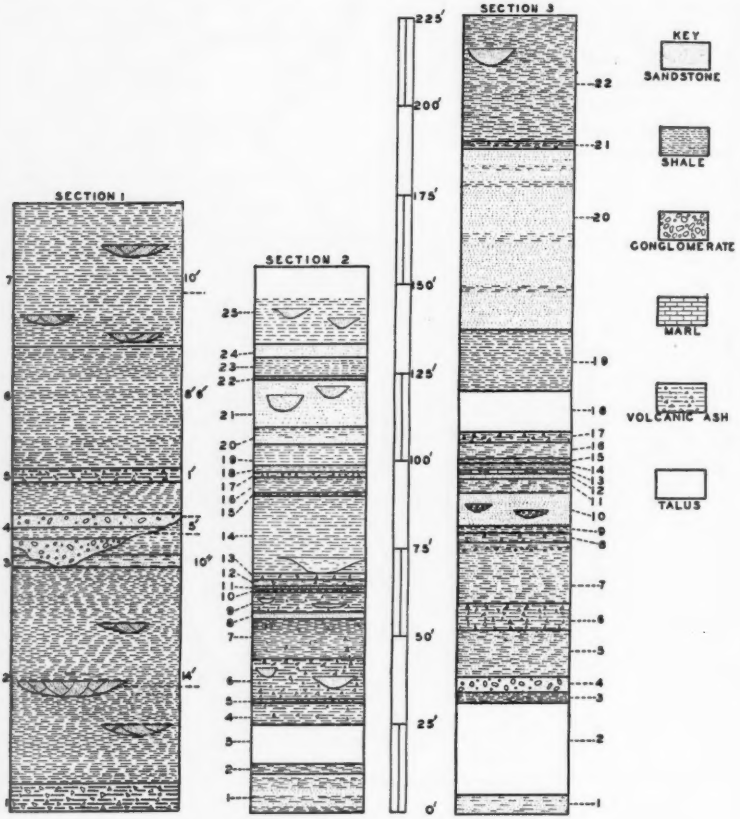


FIGURE 4.—*Measured sections of Gracias formation, lower Pliocene*

1 (Fig. 4). The dip is 7° NE. Plants and vertebrates comprise the fossils of the locality.

Section 1	Thickness Feet	Inches
7. White clay with many cross-bedded channels or lenses of sandstone. Fragments of fossils in the channels	10	
6. Light, yellow clay. No fossils	8	6
5. Fine, white, volcanic ash. No fossils	1	

	Thickness Feet	Inches
4. Flesh-colored clay cut by deep stream channel. Channel chiefly coarse gray sandstone with pebbles and small cobble scattered through it. Thin layer of clay between layers of sand and conglomerate. Sand and conglomerate containing well-preserved vertebrate fossils. Plant remains in clay bed	5	
3. Coarse brown sand with some clay. Cut by large channel of 4	10	
2. Yellowish clay with numerous cross-bedded sandstone channels, conglomeratic in places. Fragments of vertebrates in channels	14	
1. Fine, pure volcanic ash containing no fossils. Base not exposed, thickness uncertain.		

Section 2 (Fig. 4) is in the vicinity of Rancho Lobo about 1 mile north of Section 1 on the west side of the Lobo River. It gives a more complete sequence of beds but has not yielded so complete a fauna. Channel sands are not so well developed.

Section 2

	Thickness Feet	Inches
25. Shale with sandstone channels. Thickness uncertain.		
24. Fine, white, cross-bedded sandstone. No fossils	3	6
23. Brown clay. No fossils	5	0
22. Fine, pink clay. No fossils	1	0
21. Sandstone cut by cross-bedded sandstone channels. Fossils in channels.	13	6
20. Sand, grading to clay at top. No fossils	4	0
19. Light-colored clay. No fossils	5	6
18. Hard, brown clay. No fossils	1	8
17. Hard, white, volcanic ash. No fossils	1	6
16. Clay. No fossils	4	0
15. Fine, white, volcanic ash	1	0
14. Yellow clay containing large sandstone channel. Little volcanic ash. Fossils in channel	22	0
13. Hard, white volcanic ash. No fossils	2	0
12. Ashy clay. No fossils	2	0
11. Massive sandstone. No fossils	1	0
10. Hard, white, ashy clay. No fossils	1	0
9. Ashy clay with micaceous sandstone channels. No fossils	6	4
8. Fine, white, massive sandstone. Channel-like in character, cutting down to unit 7 in places. No fossils	2	0
7. Ashy clay, white. No fossils	11	
6. Hard, white, volcanic ash with small amount of clay. A few sandstone channels. Fragments of vertebrates in channels	12	
5. Soft, white, volcanic ash. No fossils		8
4. Light-brown, ashy clay. Weathering to dark brown. No fossils	6	
3. Talus-covered slope	10	
2. Sandy clay. No fossils	3	
1. Yellowish clay. Base not exposed. No fossils.		

These two sections, on the basis of their elevation and the general similarity of beds, appear to have been deposited at about the same time, but it is not possible to correlate exactly any beds or sequences of beds. Tracing is made impossible by the heavy growth of vegetation. In all probability there was little actual lateral continuity of the beds in the region. Most of the fossils occur in the channel sands, although, in a few places, fragments of teeth were found in the more characteristic layers. The plants occur in shales or clays overlain and underlain by vertebrate-bearing horizons. There can be no doubt of the contemporaneity of the fauna and flora.

Tapasuna-Mangual locality.—This locality (Fig. 3) was first reported by Frick (1929). It lies about 10 miles northwest of Gracias on the west bank of the Mejocote River. Exposures are much less well developed than in the Rancho Lobo locality, no true badlands being present. Likewise, the fauna is poorer in number of individuals and perfection of preservation. Section 3 (Fig. 4) was measured about a mile northeast of Tapasuna.

Section 3

	Thickness Feet	Inches
22. Yellow, sandy clay with sandstone channels	33	
21. Laminated, volcanic ash and clay	3	
20. Fine, yellow sandstone alternating with clay and sandy clay	50	
19. Yellow clay	17	
18. Talus-covered slope	11	
17. White, volcanic ash	3	
16. Yellow, sandy clay	5	
15. Pink clay	1	
14. Yellow sandy clay	1	
13. Pure, white marl	1	6
12. Fine, pure sandstone	10	
11. Sandy clay	4	
10. Fine, white sandstone with channels of coarse material	10	
9. Light-colored clay	2	
8. Light, ashy clay with some sand	3	
7. Sandy clay. Pebble layer near top	13	
6. Volcanic ash. Some clay	7	6
5. Grayish-yellow sandy clay	13	
4. Well-cemented conglomerate. Maximum pebble size 2″ diameter	4	
3. Brown clay		2
2. Talus-covered slope	27	
1. Sandy clay	5	6

This section is similar in general composition to those at Rancho Lobo but differs in the higher percentage of sandy clay and the lower

percentage of volcanic ash. Conditions of deposition appear to have been essentially similar in the two regions visited, but the Tapasuna-Mangual formations may have been somewhat farther from the main stream channel than those at Rancho Lobo. Exact correlation between the two areas is impossible, but the faunas suggest that the time of deposition was essentially the same.

GEOLOGICAL HISTORY OF THE REGION

The formations of the Mejocote valley, the deposition of the Tertiary beds, and the subsequent shifting and erosion of the beds present a short, incomplete but interesting picture of the conditions in this region during the late Tertiary. The valley in this region is quite certainly structural in origin. The mountain ranges to the east and west were formed in pre-Pliocene time and may have assumed the general outlines which they have today during Miocene folding. During the lower Pliocene, erosion of the highlands bordering the valley produced a great volume of sediment which was spread out over the floor of the valley on broad flood plains. Within the channels were deposited sand, gravel, and, in some cases, boulders; the rains must have been torrential at times. Mixed with the material derived from erosive processes is much volcanic ash testifying to the activity of near-by volcanoes during the lower Pliocene. Some thin beds contain as much as 90 per cent volcanic ash.

At about the end of the lower Pliocene marked changes appear to have been initiated. The streams, which had been depositing hundreds of feet of material, became sufficiently active so that erosion of the valley floor overshadowed deposition of new material. Evidence of the factors which brought about this change are not conclusive. It appears, however, that there was a marked uplift causing a rejuvenation of the streams and initiating a cycle which has continued to the present time. During this uplift, or perhaps subsequent to it, the lower Pliocene beds were tilted to various degrees and in various directions. The terracing of the streams suggests that there were actually several pulsations in the uplift, but the situation is complicated by the fact that some of the terraces are controlled by the presence of resistant strata. More detailed study is necessary to determine the relative influence of the two factors. The present condition of the valley is probably not greatly different from that which existed during the late Pliocene and the Pleistocene.

FAUNA

GENERAL STATEMENT

Most of the fauna came from three localities: the Rancho Lobo locality, the Tapasuna locality, and a locality 3 miles north of Rancho Lobo on

the Tapasuna trail. The following forms were obtained from these localities:

(1) RANCHO LOBO:

Pliohippus hondurensis n. sp. (W. M. 1769, 1770, 1771, 1772, 1773, 1774, 1776, 17777, 1778, 1779, 1780, 1781, 1782, 1783, 1789)
Neohipparion montezuma (Leidy) (W. M. 1788)
Cervid indet. (W. M. 1775)
Procamelus sp. (W. M. 1791)
Amphicyon sp. (W. M. 1767)
Blickotherium cf. *blicki* Frick (W. M. 1761, 1762)

(2) 3 MILES NORTH OF RANCHO LOBO:

Pliohippus hondurensis n. sp. (W. M. 1784)
Teleocerine rhinoceros (W. M. 1790)
Blickotherium cf. *blicki* Frick (no numbered specimens)

(3) TAPASUNA-MANGUAL:

Pliohippus hondurensis n. sp. (W. M. 1792)
Neohipparion montezuma (Leidy) (W. M. 1787)
Blickotherium cf. *blicki* Frick (W. M. 1763, 1764)

In addition a few indeterminate fragments of rodents were found, as well as some indeterminate postcranial remains, parts of skulls, and poorly preserved teeth. Mastodon remains were found in many localities, but for the most part identification is impossible. Much of the larger mastodon postcranial material was not carried out because of its poor state of preservation and the necessity of shipping all specimens by airplane.

DESCRIPTION OF THE FAUNA

Order PERISSODACTYLA

Family EQUIDAE

Pliohippus hondurensis n. sp. (Pl. 1, A-D; Pl. 2, A-D, F-H; Pl. 3, A-F, Fig. 5).

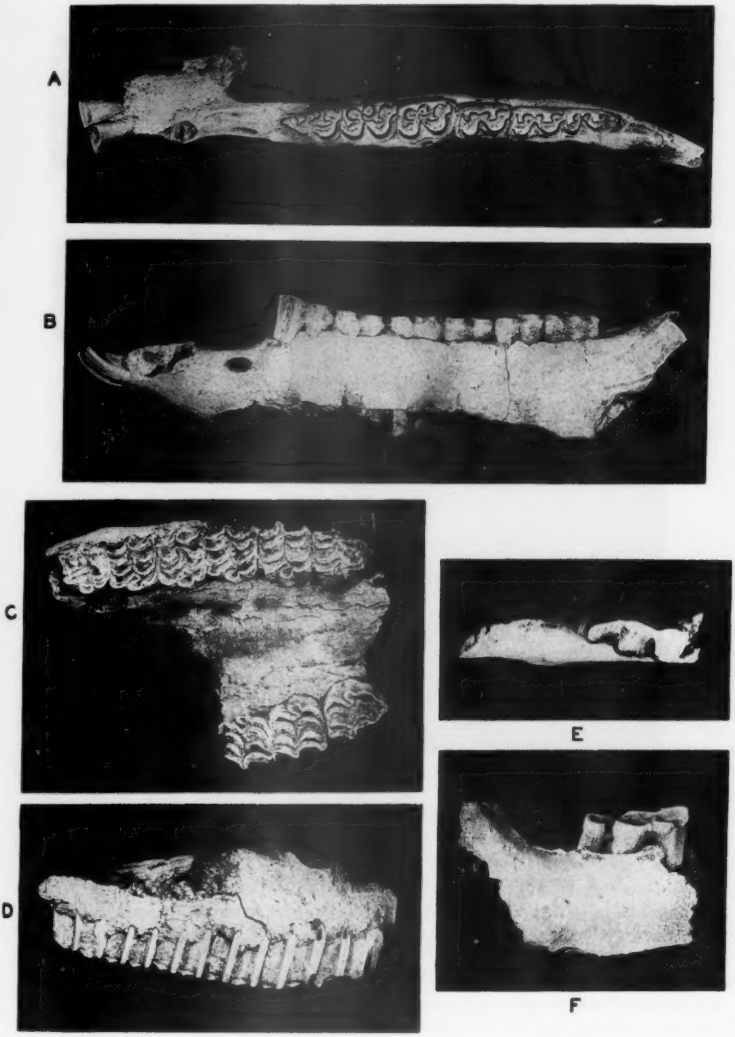
HOLOTYPE: W. M. 1769. Partial skull containing left P^2—M^3 and right P^2—^{4}.

HORIZON AND LOCALITY: Gracias formation, lower Pliocene. Rancho Lobo, 3 miles north of Garcias, Honduras.

PARATYPES: W. M. 1770, pair of lower jaws with right P_3—M_3.
W. M. 1771, pair of lower jaws with right dentition complete except for $I_{\frac{1}{1}}$, left dentition complete except for I_1 and M_3.
W. M. 1772, maxillary with P^2—M^2.
W. M. 1773, lower jaws with I_1 and P_3—M_3.

DIAGNOSIS: *Pliohippus hondurensis* may be distinguished by its small size (*See* Measurements) and relatively small protocone.

DESCRIPTION: The occlusal surface of the cheek teeth of *P. hondurensis* is one-third smaller than those of the smaller plains species of *Pliohippus*, whereas the crown is approximately of the same length. The upper cheek teeth are strongly curved. The crown pattern is simple, the unworn teeth being marked only with a few minor plications which disappear with wear. In unworn, or little worn teeth, the following plications may be observed: Two or three minor plications on the anterior enamel border of the postfossette (pli-post-fossette, Osborn); two deep plications on the posterior of the prefossette (pli-pre-fossette and pli-protoconule, Osborn); and, an inner plication on the crochette between the metaloph and protoloph (pli-caballin, Osborn). These plications are quite variable in size and in some individuals carry



PLIOHIPPUS HONDURENSIS n. sp. and INDETERMINATE CERVID

A. *P. hondurensis* n. sp., paratype, W. M. 1773, dorsal view, x ½.

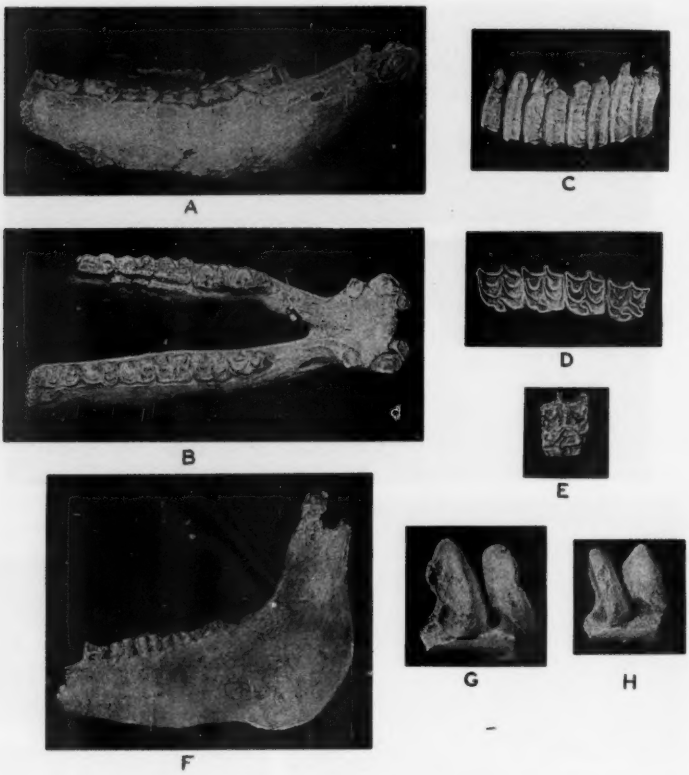
B. W. M. 1773, lateral view, x ½.

C. *P. hondurensis* n. sp., holotype, W. M. 1769, ventral view, x ½.

D. W. M. 1769, lateral view, x ½.

E. Cervid, indeterminate. W. M. 1775, dorsal view, x 1.

F. W. M. 1775, lateral view, x 1.



PLIOHIPPUS HONDURENSIS n. sp. and *NEOHIPPARION MONTEZUMA* (Leidy)

A. *P. hondurensis* n. sp., paratype, W. M. 1771, lateral view, x 3/8.

B. W. M. 1771, dorsal view, x 3/8.

C. *P. hondurensis* n. sp., W. M. 1774, lateral view, x 1/2.

D. W. M. 1774, dorsal view, x 1/2.

E. *N. montezuma* (Leidy), W. M. 1787, M^1, crown view, x 1/2.

F. *P. hondurensis* n. sp., W. M. 1770, lateral view, x 1/4.

G and H. *P. hondurensis* n. sp., astraguli, x 1/2.

farther toward the bases of the teeth than in others, The hypoconal groove (between the hypocone and hypostyle) is variable in form and depth. In most individuals it is open to within a quarter of an inch of the base of the tooth. At this level it merely flattens out. In some teeth it becomes closed at about half the distance to the base of the tooth and forms an isolated lake.

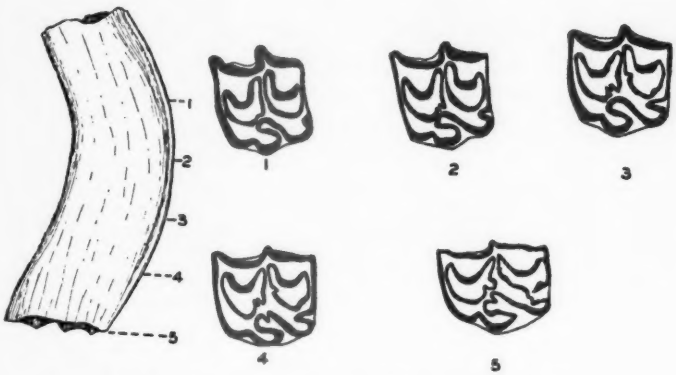


FIGURE 5.—*Enamel pattern of M¹ of* Pliohippus hondurensis n. sp.
At successive stages of wear (×½).

The protocone is proportionately small and is invariably connected to the protoconule by a rather narrow isthmus. In little worn teeth the protocone is elongated, terminating posteriorly in a sharp point. With a little wear the protocone becomes almost round and remains so almost to the base of the tooth. In extremely heavily worn teeth the protocone may become connected with the hypocone, leaving a small enamel lake external to the point of union. The hypocone is slightly smaller than the protocone. In little worn teeth it is distinct and similar in shape to the protocone. With wear, however, when the hypocone groove becomes shallower and finally disappears, the hypocone unites with the hypostyle and loses its round shape. The parastyle and mesostyle are normally heavy and prominent whereas the metastyle is reduced. Cement fills the fossettes in all cases. On the outer and inner surfaces of the teeth the amount of cement varies from thick to complete absence. It is possible that it was actually more uniform when the animals lived and the minor differences in the conditions of preservation cause the variations observed in the fossils.

The lower cheek teeth are relatively long crowned and taper gently to their bases. The premolars are slightly longer antero-posteriorly than the molars and are much wider transversely. The metaconid and metastylid of the first premolar are very small and strongly united. The metaconids and metastylids on the third and fourth premolars are large, round, and united by a narrow isthmus. The V's formed internally by the enamel borders between the metaconulids and metastylids are rather deep in unworn posterior premolars. When the teeth are heavily worn this V disappears, and the inner border is only slightly concave. In the premolars the enamel fold between the metastylid and hypoconid runs far forward terminating at a point

bucal, or slightly antero-internal, to the gutter between the protoconid and hypoconid. The molar pattern is quite different from that of the premolars. The metaconid and metastylid are very small, and the inner groove, between them, is only slightly concave even in nearly unworn teeth. The enamel fold between the metastylid and entoconid is shallow and does not extend anteriorly as it does in the premolars.

There is considerable variation in the general proportions and shape of the ramus of the jaw of *P. hondurensis*. These differences are probably due to differences in age and sex and to individual variation. The ramus of a fairly young, but mature individual (W. M. 1773, Pl. 1, A and B) has a slender, long, almost horizontal symphisial region with procumbent incisors. An older individual (W. M. 1771, Pl. 2, A and B) has a symphisial region which is more robust, heavy, wide, and ascending at a much greater angle with the incisors much less procumbent. The diastema between the canine and second premolar is longer in the older individual. The mental foramen, lying below and a little anterior to the second premolar in both specimens, is round and large in W. M. 1773 and elongated and small in W. M. 1771. The angle and ascending ramus are like those in other species of *Pliohippus*, the angle being deep and the ascending ramus tapering gradually upwards.

The small size of *P. hondurensis* recalls the progressive reduction of tooth size in *Nannippus*. The Honduras species cannot be referred to *Nannippus* because of its consistently connected protocone.

"Protohippus" placidus[3] resembles *Pliohippus hondurensis* in its small size and connected protocone. The straight cheek teeth and simple pattern of *"Protohippus" placidus* bar it from any close relationship to *Pliohippus hondurensis*.

The species which appears to be most closely related to *P. hondurensis* is *P. castilli* Cope. This species comes from the lower Pliocene of Vera Cruz, Mexico. It is small and differs from the Honduras specimens mainly in the small size of its hypocone and the relatively large protocone. It is probably significant that the two smallest species of *Pliohippus* occur in Mexico and Honduras. The possible implications of this fact are discussed in a subsequent section of this paper.

There is nothing in the structure of *Pliohippus hondurensis* which suggests an approach to the *Hippidium* group of South America.

Tooth Measurements: Tables 1-3 have been compiled from measurements taken, except in cases of measurements of crown length, from complete or relatively complete series of dentition. Measurements of individual, isolated teeth indicate that size variation in the species is slightly greater than that shown in the tables. Measurements of the antero-posterior (a. p.) and transverse (tr.) dimensions are based on the limits of the enamel since the cement is somewhat variable in its development. This accounts for the discrepancies between the length of the tooth row as indicated in Table 1 and the cumulative length which may be obtained by taking the sum of individual measurements. The crown length measurements have been made on the few available unworn molar teeth. The upper crown length was measured along the enamel surface of the mesostyle, and the lower crown length along the enamel surface of the protoconid.

In addition to teeth and jaws, numerous parts of the skeleton of *P. hondurensis* were recovered including two tibiæ (W. M. 1780), several astragali (W. M. 1779), two pelves (W. M. 1778), lateral metapodials (W. M. 1782), and various phalangeal elements and vertebræ (Pl. 2, G and H; Pl. 3, A-F). The lateral metapodials are

[3] This species certainly is not congeneric with species originally described as *Protohippus perditus* (McGrew and Meade, 1938; Stirton, 1940). *Protohippus* is synonymous with *Merychippus*.

PLIOHIPPUS HONDURENSIS n. sp.

A and B. Tibiae, anterior and posterior views, x 1/2.

C. W. M. 1786, lateral metapodial, x 2/3.

D and E. Phalanges, x 2/3.

F. W. M. 1786, medial metapodial, x 2/3.

A

E

B

F

C

G

D

BLICKOTHERIUM, AMPHICYON, AND *PROCAMELUS*

A. *B.* cf. *blicki* Frick, W. M. 1761, right M^3, x 1/3.

B. *B.* cf. *blicki* Frick, W. M. 1762, left M^3, x 1/3.

C. and D. *B.* cf. *blicki* Frick, partial tusks, C showing enamel band. C, x 1/4. D, x 1/3.

E and F. *Amphicyon* sp. W. M. 1766, ramus of jaw. E, medial, F, lateral. x 1/3.

G. *Procamelus* sp., W. M. 1768, distal end of tibia, x 1/3.

TABLE 1.—*Measurements of crown length and length of tooth row*

W.M. No.	Unworn crown length, mm.		Length of tooth row, mm.				Fraction of wear of crown length
	M^1	M_1	P^1-M^1	P^2-M^2	P^4-M^4	P_1-M_1	
1782	55.0	46.0	.....	.....	.....	.....	unworn
1796	.....	.....	112.20	96.0	64.5	.....	1/2
1772	.....	.....	.....	99.5	.....	.....	1/5
1774	.....	.....	.....	.....	66.0	.....	3/5
1773	.....	.....	.....	.....	.....	104.0	1/4
1770	.....	.....	.....	.....	.....	99.0	2/5
1777	.....	.....	.....	.....	.....	102.0	1/2
1781	.....	.....	.....	.....	.....	101.0	1/2

TABLE 2.—*Individual upper cheek teeth*

W.M. No.	P^2		P^3		P^4		M^1		M^2		M^3	
	a.p.	tr.	a.p.	tr.	a.p.	tr.	a.p.	tr.	a.p.	tr.	a.p.	tr.
1769	20.5	15.1	17.1	18.5	16.6	18.2	15.2	18.0	15.1	17.0	15.2	15.2
1772	23.0	17.8	19.0	19.5	18.8	19.8	17.0	20.0	17.0	18.0	....	....
1774	....	....	....	....	18.2	19.0	16.0	18.8	15.0	18.2	15.0	16.0

TABLE 3.—*Individual lower cheek teeth*

W.M. No.	P_2		P_3		P_4		M_1		M_2		M_3	
	a.p.	tr.	a.p.	tr.	a.p.	tr.	a.p.	tr.	a.p.	tr.	a.p.	tr.
1771	18.8	10.0	16.8	10.8	16.5	9.9	15.2	8.2	15.6	8.0	20.5	....
1773	18.9	9.5	17.2	11.0	17.0	10.0	16.1	8.7	16.6	7.9	18.1	6.8
1777	18.0	9.8	17.1	10.1	16.2	9.8	15.1	8.2	15.8	7.5	20.0	7.2
1881	....	....	10.0	11.1	17.5	10.1	15.2	9.2	....	....	....	....

proportionately much the same as those of the larger plains species of *Pliohippus* of North America and indicate a similarity of foot structure between the Honduras and North American forms. The phalangeal elements are slender as in other species of *Pliohippus*. The pelves are relatively very lightly constructed. Except for size and general lightness of structure *P. hondurensis* cannot be distinguished from other species of *Pliohippus* on the basis of postcranial material.

Neohipparion montezuma (Leidy) (W. M. 1787, 1788)

(Pl. 2, E)

Several teeth from the Tapasuna locality (W. M. 1787) and a poorly preserved maxillary with partial dentition from a channel of Rancho Lobo (W. M. 1788) are referred to this species.

The tooth measurements are as follows:

	M^1	P^2
Specimen	W.M. 1787	W.M. 1788
Antero-posterior, crown	16	25
Transverse, crown	19	18
Antero-posterior, protocone	8	6
Transverse, protocone	4	indet.

M^1 is quite worn, and it is not possible to determine the complete crown length. The small size, the elongated protocone, and the nature of the plications are the principal characters upon which specific reference has been based. In the case of W. M. 1788 from Rancho Lobo, the poor condition of the dentition makes specific determination less certain.

Family RHINOCEROTIDAE

Teleocerine rhinoceros (W. M. 1790)

Only fragmentary remains of rhinoceroses were found in the Gracias formation. A partial ramus of a jaw and a single, complete cervical vertebra comprise the complete list of specimens. It is not possible to do more than place these specimens in the general teleocerine group. The specimens were obtained from the Tapasuna locality.

Order CARNIVORA

Family CANIDAE

Amphicyon sp. (W. M. 1766, 1767)

(Pl. 4, E and F)

This genus is represented by two specimens, one a mandible from Tapasuna (W. M. 1766), and a second, a single M^1 from Rancho Lobo (W. M. 1676). Neither of these possesses characters which permit a certain specific assignment. In the lower jaw only the heavy canine tooth is present. The ramus is heavy but tapers anteriorly from the posterior angle so that it is only two-thirds as deep under the symphysis as under the alveolus of M^2. As preserved, the ramus measures 160 mm., but the complete length was about 175 to 180 mm. The jaw is typically amphicyonine, but more accurate placement depends upon the finding of more adequately preserved material.

The single M^1 which was found at Rancho Lobo lacks part of the paracone and metacone and is considerably worn. It is clearly referable to *Amphicyon* and is a member of a relatively small species.

Order ARTIODACTYLA

Cervid Indeterminate (W. M. 1775)

(Pl. 1, E)

A portion of a ramus containing a heavily worn M_3 represents a small pecoran, probably a cervid, which was found near Tapasuna. In size it is equal to *Pudu* or *Mazama*. The molar is subhypsodont. The external groove between the first and second lobes extends a little more than half way across the tooth, even in its very worn condition. The third lobe is relatively narrow transversely, and the antero-posterior length is but slightly less than that of the first and second lobes. The enamel border of the antero-external portion of the second lobe is extended indicating that a basal tubercle was present. Posterior to M_3, the ramus has a high internal crest and slopes gradually externally; there is no hollow such as is found in most cervids.

In general, this specimen appears to resemble *Aletomeryx* in the hypsodonty of the molar and in the lack of a hollow behind M_3. It differs from that genus in having a much deeper V between the protoconid and the hypoconid in the heavily worn M_3. It also has a strong resemblance to *Mazama* in size and proportion, but M_3 is much more hypsodont. Closer definition of the affinities of this form will be dependent upon the discovery of more complete specimens.

Family CAMELIDAE

Procamelus sp. (W. M. 1768, 1791)

(Pl. 4, G)

The camels are represented by three specimens, two distal ends of tibiae from Tapasuna (W. M. 1768) and a partial radius from Rancho Lobo (W. M. 1791). In size these elements are comparable to those of large specimens of *Procamelus* from the Great Plains of North America. There are no characters present which indicate any generic distinction from *Procamelus* and none which allow specific assignment.

Order PROBOSCIDEA

Family BUNOMASTODONTIDAE

Subfamily RHYNCHOROSTRINAE

Blickotherium cf. *blicki* Frick (W. M. 1761-1764)

(Pl. 4, A-D)

Most, if not all, of the proboscidean material collected may be referred with little question to the genus *Blickotherium* and tentatively to the species *blicki*. Although a great many fragments of mastodon remains were encountered only a few specimens were collected because the state of preservation of most of the material did not seem to justify the expense of shipping. Three cheek teeth which were collected do not differ in any major feature from those of *Blickotherium blicki* and may be referred tentatively to that species. No description is necessary since the characters are shown in the figures.

Several sections of incisor tusks quite certainly belong to this genus and species. They have a characteristic lateral compression and a well-developed enamel band. One partial tusk (W. M. 1764) does not have an enamel band developed and for this reason might be referred to the genus *Aybelodon*. Such an assignment is rather dubious, however, since it is possible that the band was lost in the course of

fossilization or, if this represents an old specimen, was lost with advancing age. This section of tusk is less compressed laterally than the others. Although it may belong to a separate genus, there is no conclusive evidence that it is to be excluded from *Blickotherium blicki*.

FLORA

Associated with the vertebrate remains in many places were pieces of fossil wood and leaves of palms and grasses. W. C. Darrah of Harvard University has identified the woods and has offered advice as to the possible significance of the genera which occur. The writers wish to express their gratitude to him for this work which has been of great value.

The flora list is small, certainly very incomplete, but it is of considerable significance in the determination of the environment in which the fauna lived. Specific identifications of the wood specimens were impossible. The palms and grasses have not been identified and occur in such a fragile matrix that detailed work on them is next to impossible. The floral list is as follows:

Family SAPINDACEAE *Schmidelia* sp. indet.
Family CELASTRACEAE *Gyminda* sp. indet.
Legume Genus indet.
Palms Genus indet.

Only the more resistant structure of the woods are evident, and this, along with the difficulty of determining species in fossil wood when well preserved, has made it unwise to attempt any specific assignment. The cell walls of the legume are black, while those of the other woods are rather clear and colorless.

Schmidelia and *Gyminda* are typical of hardwood rain forests and occur at present in Central America, northern South America, and the Caribbean Islands, extending northward into southern Mexico. The trees also occur in broken forests on the margins of the rain forests, and it is entirely possible that they may have been associated with the palms and grasses which occur in essentially the same deposits. The field evidence is not sufficient to assure that such an association actually existed. Much of the wood is in small pieces, and all of it occurs in sediments which have undergone some transportation before deposition. It is probable that the environment indicated by the woods and the palms is that in which the fauna lived, but, while we may be certain of association of the palms, it is not impossible that the wood actually grew some distance away. In general, it is safe to say that the environment was not greatly different from that represented by the hardwood rain forests or the broken periphery of these forests today and that conditions were quite different than those on the great plains to the north at about the same time.

The plants do not add to our data for determination of the age of the Gracias formation. *Schmidelia* and *Gyminda* are known in various regions of Central and northern South America from Eocene to recent time.

DISCUSSION

FAUNAL AND FLORAL ASSOCIATIONS

The vertebrate fauna from Honduras contains elements which commonly inhabit the plains, notably the horses, and others which are more characteristic of forested areas—the rhinoceros, deer, and proboscideans. Those forms which characteristically occur in forest regions are equal in size to their northern contemporaries, while the horses are much smaller. The fragmentary remains of carnivores indicate animals no smaller than similar forms in the Pliocene of the United States. The association of horses with palms and wood from trees of tropical or semitropical rain forests suggests that their small size is due, in part at least, to their existence in an environment to which their structure is not particularly suited. The other forms show no effects of the differences in climate which existed between the areas in which they occurred. Although the horses were in what would appear to be an unfavorable environment, they thrived as witnessed by the fact that they are by far the most numerous animals present.

CORRELATION

Accurate correlation of the Honduras fauna with better-known faunas is rendered difficult by several factors. The distance and the latitudinal difference of the fossil-bearing deposits from any other closely studied horizons of approximate contemporaneity give rise to a number of problems. Comparisons are necessarily made with specimens from the Great Plains and from California where the mammals, especially the horses, have been carefully studied and their succession well worked out. The environmental difference which must have existed between Honduras and the western United States is an important factor which seems to have had more effect on the horses than on the other elements of the fauna. Matthew's hypothesis that homotaxis does not imply synchrony when dealing with faunas widely separated latitudinally, especially when one is in the Tropics and the other in the Temperate Zone, must be considered.

Fortunately horses are well represented and, as usual, form the most reliable basis for correlation. The structure of the Honduras teeth indicates a stage of progression equivalent to that reached by lower Pliocene horses of the Great Plains. The significant characters, hypsodonty and crown pattern, may be nearly matched in specimens from Clarendon.

Burge, and Ricardo faunas. Variations in tooth sizes was independent of change in these characters and the extremely small size of the teeth in *P. hondurensis* is of no value in time correlation. Homotaxially the horses of the Gracias formation are equivalent to the Clarendonian horses of the United States.

According to Matthew's hypothesis of mammalian dispersal a fauna from Honduras should have lived considerably later than its homotaxial equivalent in the Nearctic realm. On the basis of this reasoning, the Honduras fauna, being homotaxially about equivalent to the Clarendonian fauna of the Great Plains, should be at least as recent as Hemphillian. In cases where it is evident that mammalian dispersal is dependent upon climatic changes, the validity of Matthew's arguments hardly can be doubted, but in this case, where it seems reasonably certain that the fauna expanded its range from a favorable habitat to one less favorable, these arguments cannot apply.

There can be little doubt that the plains offered the best ecological conditions for the existence and development of an animal with the particular structural features of the horse. The spread of the horse into tropical, or subtropical, forested, and probably mountainous regions of Central America must have represented a profound change of habitat and it is a clear case of migration of a stock into an environment less favorable for its existence. Any one of a number of factors might be responsible for such a migration. Radiations into peripheral zones will tend to occur from any well-populated area provided the conditions in the adjacent zones are not such as to prohibit such movements. Overpopulation of the plains would hasten such a process. That the range of *Neohipparion* and *Pliohippus* was continuous from the United States southward through Mexico, Guatemala, Honduras, and probably to the very edge of the Pliocene Central America peninsula is indicated by the occurrence of *Neohipparion peninsulatum, Neohipparion montezuma,* and *Pliohippus castilli* in Mexico. These species are small and in other respects similar to the horses from Honduras. It is probable that there was a continuous range of horses from the plains of the United States to Central America with a gradual diminution of size to the south. Bergmann's principle that closely related, warm-blooded animals tend to be smaller in warm than in cold climates may explain the small size of the Honduras horses. The principle cannot be applied to the other constituents of the Gracias fauna since they seem to be no smaller than their northern relatives. It is very probable that the small size of the horses in the south is the result of penetration into an unfavorable environment whereas the same was probably not true of the rhinoceros or deer.

If the above interpretation of the homotaxial equivalency and modes of migration be correct, there is no reason to believe that the Gracias fauna is any younger than Clarendonian. It is on the basis of these interpretations that the age of the formation has been determined as lower Pliocene.

LAND CONNECTIONS

The present work offers some additional evidence bearing upon land connections between the American continents and between North America and Central America. Considering first the connections between Mexico and Central America, it is apparent that the vertebrate fauna has an interesting bearing on the time of submergence of the isthmus of Tehauntepec which lies between the Gulf of Mexico and the Gulf of Tehuantepec of the Pacific. It has been conceded by most geologists and paleontologists working on material from this area that the thick limestones in the mountain zones and the presence of Atlantic faunal elements in the Pacific region are good evidence of a marine connection across the isthmus during middle or late Tertiary times. The exact time of the development of the portal has not been certainly determined. Spencer (1897), Vaughan (1919), and a number of others, constituting a majority of those who have worked in the region, concluded that the date of inundation was very late Miocene or early Pliocene. Others, particularly Dickerson (1917) and Woodring (1931), concluded that the portal was open before the late Miocene.

If Central America was isolated during the lower Pliocene it would be very difficult to account for the Honduras mammalian fauna. Two possibilities, however improbable, present themselves: (1) The fauna is lower Pliocene in age and was derived from one present in the region during the Miocene. This would automatically imply the independent development of *Pliohippus* and *Neohipparion* from some Miocene ancestor. That such is true seems unlikely. Also it is doubtful if any proboscideans existed sufficiently early in the Miocene in Honduras to give rise to the abundant members of this order in the fauna; (2) the second possibility is that the fauna is of middle Pliocene age, having migrated from the north after closure of the portal. This, for reasons stated above under the heading Correlation, does not seem to be the most reasonable conclusion.

The conclusion that seems most likely is that the route to Central America was open during the lower Pliocene and that the Honduras fauna was continuous with that of North America. This would support the conclusion of Woodring, Dickerson, and others that the portal was not open after the close of the Miocene.

Some new evidence is also provided relating to the Central America–South America land bridge. It is generally accepted (although not uni-

versally) that a through migratory route between North and South America was not opened until very late Pliocene times. Schuchert (1935, p. 557) suggested that the lower Pliocene barrier was at the Tehuantepec instead of the Panama isthmus.

> "As the Tehuantepec portal of southern Mexico was the last to close (in middle Pliocene at the latest), the South American mammals could have spread into Central America after the older portals of the Panama and Costa Rica had been transformed into land (late Miocene), but none could have gotten north of Tehuantepec until after the middle Pliocene, when this portal was again changed to dry land."

As shown above, the lower Pliocene fauna in Honduras almost conclusively proves that the Tehuantepec portal was not open during the lower Pliocene, hence the only barrier between North and South America was at the Panama isthmus.

The fact that no mammals of southern origin were encountered in the lower Pliocene of Honduras strongly supports the growing belief that no land connection was available between North and South America until late Pliocene. The megalonychid sloths found in several lower Pliocene localities in North America and the procyonids and didelphines which appeared in the lower Pliocene (or upper Miocene) of South America were apparently "barrier jumpers" and did not reach their new habitat by a dry land route (Patterson, 1937; Simpson, 1940).

It has been suggested that the barrier, in a large part, was of an environmental nature, that tropical rain forests prohibited the exchange of faunas between the two American continents (Berry, 1918, p. 636). The presence of this typically North American fauna in a presumably tropical climate, in or near the hardwood rain forests, and in country which certainly had considerable relief, implies strongly that such factors did not form an effective barrier at least in post-Miocene times. Certainly some mammals would be stopped by such conditions but not enough to make any difference in the general picture. For example: The late Pliocene influx of northern mammals into South America is very definite although nearly half of the North American mammalian families failed to migrate to the southern continent (Simpson, 1940), presumably because of environmental inclemency in Central America.

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FLOW CLEAVAGE IN FOLDED BEDS

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ABSTRACT

Flow cleavage is distinguished from certain allied structures by its lack of banding and the characteristic development of new minerals that exhibit parallel crystallographic orientation. These petrographic features show that flow cleavage is due to recrystallization under differential stress rather than to orienting mechanical movements. Supporting this conclusion are the relations shown by simple folds in originally sandy and shaly beds, the argument being that the deformation of the slaty beds could not create orienting movements of such character and intensity as to produce flow cleavage which is uniform across the crests of the folds and is well developed even in open folds. In field work where the relations between bedding and flow cleavage or allied structures are used to determine the character of the major folds, the principle that flow cleavage is parallel to the axial planes of the folds is of proven value. Field relations do not accord with certain other principles that have been advocated, such as the principle that foliation is parallel to the axial plane only in slip folds or the principle that axial plane foliation can be distinguished from bedding foliation by means of certain criteria derived from a statistical analysis of the fabric. In complex structures the relations between flow cleavage and bedding may not everywhere agree with the principle stated above. However, where abnormal relations occur they are believed to be due to complicating factors such as the presence of two periods of deformation, and several examples are given to support this viewpoint.

INTRODUCTION

Originally this paper was prepared for the annual meeting of the Geological Society in December 1939, as part of a program devoted to the discussion of the plastic deformation of the solid earth. The present writer was asked to treat some phase of the subject from the viewpoint of the field observer and choose as his topic the relations shown by flow cleavage in folded beds. In presenting the paper at the meeting, the writer confined his remarks rather strictly to this topic. However, certain comments that have been offered since that time suggest that many readers will have in mind not only flow cleavage proper but also other structures such as gneissic banding. For this reason the treatment in the present paper has been made somewhat broader than it was in the original manuscript.

DEFINITIONS OF FLOW CLEAVAGE AND ALLIED STRUCTURES

Deformed rocks show several types of secondary structures to which various names have been applied. In the writer's opinion it is important to distinguish certain of these structures from one another and it is especially desirable to note certain fundamental differences between flow cleavage, gneissic structure, fracture cleavage, and cataclastic structure.

FLOW CLEAVAGE

According to Leith (1923, p. 113):

> "Flow cleavage is a structure commonly resulting from the flowage of hard rocks. It is a capacity to part along parallel surfaces determined by the parallel arrangement of the longer axes of unequidimensional mineral particles and by the parallel arrangement of mineral cleavage in certain of the unit mineral particles. Flow cleavage is characterized by platy and columnar minerals of comparatively few kinds which are well adapted to conditions of rock flowage. . . . Other names for flow cleavage are *schistosity* and *slaty cleavage*. . . . Slaty cleavage is ordinarily distinguished from schistosity in straightness of cleavage planes and fineness of grain, but the distinction between the two structures is not sharp—one grades into the other."

Harker (1932) gives essentially the same definitions of schistosity and slaty cleavage, and many other authors, including the writer, follow these authorities. To round out the description of flow cleavage, it may be added that minerals such as mica, chlorite, and hornblende are especially abundant and well oriented in rocks with this structure. They are largely of new crystallization, that is, they are developed at the same time as the flow cleavage, and the perfection of the flow cleavage depends mainly on their abundance. Arrangement into bands differing from one another in composition is typically absent, or, if present, it is entirely overshadowed by the common orientation of similar minerals in the various bands.

GNEISSIC STRUCTURE

Harker (1932, p. 199) says regarding gneissic structure:

"The term may be taken to imply a relatively coarse grain, with at most a very imperfect orientation of the component crystals, but often showing none the less a parallel structure of a different order (foliation)."

Regarding foliation, Harker (1932, p. 203) writes:

"Foliation consists in a more or less pronounced aggregation of particular constituent minerals of the metamorphosed rock into lenticles or streaks or inconstant bands."

Leith (1923) and many other authors, including the writer, adopt essentially the same definition of gneissic structure. The writer would emphasize, as the dominant characteristic of gneissic structure, the arrangement of the material into bands or lenses differing from one another in composition. The individual layers may or may not have flow cleavage. Commonly some do and others do not. The texture is usually medium- to coarse-grained, but the difference between a schist and a gneiss is not fundamentally a matter of grain size.

FRACTURE CLEAVAGE

Fracture cleavage is also known by various other names, including false cleavage, close-joints cleavage, and strain-slip cleavage. Insofar as the writer knows, there is general agreement on the definitions of these terms. The structure is due to the presence of closely spaced planes of parting at a certain small distance apart. The fissility is due to shear planes developed by deformation, and usually there is evidence that slipping has occurred along these planes. For example, in slates they may follow the short limbs of minute drag folds in the flow cleavage lamellae, and this condition passes into one which may be described as imbricate faulting on a very small scale. In some cases fracture cleavage closely resembles flow cleavage, but as a rule it is possible to see that the rock between the planes of parting in fracture cleavage has no structure parallel to them, or at most any parallel structure is confined to a thin film along the parting planes—that is, fracture cleavage, unlike flow cleavage, is not a property of every plane parallel to the plane of parting.

CATACLASTIC STRUCTURE

According to Holmes (1920, p. 54) the term cataclastic is:

"Applied to the structures produced in a rock by the action of severe mechanical stress during dynamic metamorphism, characteristic features being the deformation and granulation of the minerals. The term is also applied to rocks characterized by such structures."

In more detail, the deformation (without rupture) of the minerals produces features such as: secondary twinning and gliding in certain

minerals, for example calcite and plagioclase; bending of crystals, for example mica and chlorite; and permanent strain, for example that indicated by strain shadows in quartz. The granulation of the rock consists of slicing and fragmentation which starts along grain boundaries, shear planes, and so forth. On a large scale this may be called brecciation, which produces such forms as crush breccias, autoclastic rocks, and others. On a small scale, individual grains are involved, and the movement develops such well-known structures as mortar-structure, augen-structure, and mylonitic structure.

Some recrystallization naturally accompanies the deformation and granulation of the mineral grains, but its effects are subordinate to the general reduction of grain size caused by the fragmentation. However, although typical flow cleavage and typical cataclastic structure differ greatly from one another, the two structures are associated in origin and in places grade into one another. In borderline cases, the writer believes that the diagnostic feature of flow cleavage, which can be used in distinguishing it from cataclastic structure, is the development of new minerals of platy or columnar habit. On this basis, the fissility shown by some deformed limestone beds, which is clearly due to twinning, gliding, and other mechanical processes, would not be called flow cleavage. In this connection, the writer suggests that deformed rocks whose textural features are dominantly cataclastic be called sheared rocks, rather than schists, slates, or gneisses. In the case mentioned above, the deformed rock would be called a sheared limestone and not a schist.

OTHER DEFINITIONS

Some authors make no sharp distinctions between flow cleavage, gneissic structure, and cataclastic structure. For example, Knopf and Ingerson (1938, p. 7) state:

> "Foliation is used here to denote a parallel arrangement of minerals. A foliated rock will split more or less readily into thin slabs parallel to the foliation plane. Where the rock splits into thin layers, it is called a schist. Where the layers are thicker, the rock is known as a gneiss, a usage quite different from the German meaning of the word gneiss. A slate is an extremely fine foliate."

Fairbairn (1935, p. 592) concurs by saying:

> "All types of parallel texture from the finest slate to the coarsest gneiss are included under foliation."

HYPOTHESES CONCERNING ORIGIN OF FLOW CLEAVAGE

GENERAL HYPOTHESES

Various hypotheses have been proposed to explain the development of flow cleavage and allied structures. They can be briefly summarized as follows: (1) Flattening of grains or portions of the rock mass by

differential pressure; (2) solution at points where the differential stress is greatest, and deposition where it is least; (3) growth of new minerals along planes of weakness, such as joints or bedding planes, the new minerals being oriented with respect to these planes; (4) growth of new minerals along planes where the supply of material is abundant, such as bedding planes or joints serving as channelways for solutions, either aqueous or aqueo-igneous, the new minerals again being oriented with respect to these planes; (5) rotation of elongated or platy minerals into a common plane by movement during deformation, a process such as that which develops certain igneous flow structures; (6) gliding of mineral grains along preferred crystallographic directions in the mineral and along one or more shear planes in the rock, a process similar to that which develops oriented fabrics in the plastic deformation of some metals; and (7) growth of new minerals in such a way that their longest dimensions and cleavages lie in planes normal to the direction of greatest stress.

Probably none of the processes listed above can be eliminated from consideration if all structures produced by deformation of solid rocks are included. This paper, however, deals mainly with the origin of flow cleavage, and the allied structures are discussed only where such discussion seems desirable in the interest of avoiding possible misunderstandings.

As explanations of the development of typical flow cleavage, the first two hypotheses can be eliminated, because they could yield only a dimensional arrangement, that is, they would not produce the crystallographic orientation characteristic of flow cleavage. The third and fourth hypotheses can explain certain banded, that is, gneissic structures, but not typical flow cleavage. Although it is conceivable that some structures so produced may resemble flow cleavage, it is not credible that the processes called on by these hypotheses could develop typical flow cleavage, because of its lack of banding. The fifth and sixth hypotheses can be grouped together and, for brevity, called the movement hypothesis. Similarly the last hypothesis can be called the recrystallization hypothesis. These now rank as the leading hypotheses for explaining the development of flow cleavage and as such they merit a more detailed discussion.

RECRYSTALLIZATION HYPOTHESIS

The recrystallization hypothesis has been advocated by numerous authors. Leith (1923) and Harker (1932) can be named as leading authorities who favor this hypothesis.

According to this hypothesis, the systematic orientation of the mineral grains is due to recrystallization under differential stress. The characteristic minerals, such as mica and chlorite, grow in such a way that their

longer dimensions and cleavages tend to be normal to the axis of greatest stress.

The following three paragraphs, which paraphrase statements made by Harker (1932), develop this concept in more detail; also, they describe the genetic relations which, according to the recrystallization hypothesis, exist between flow cleavage and certain allied structures.

The key to the interpretation of metamorphic recrystallization is to regard it as a contrivance for minimizing the internal stresses which are in existence during the recrystallization. In accordance with this principle, the presence of a large differential stress causes tabular or elongated grains of certain minerals to form in planes normal to the greatest stress. This relation, aided by the fact that the minerals have cleavages parallel to their larger faces, causes flow cleavage. The presence of shearing stress also affects, adversely or favorably, the stability of different minerals, and we may recognize a group of minerals, called stress minerals, whose crystallization seems to be favored by shearing stress.

Mechanical processes, such as slicing, granulation, and gliding, contribute toward the development of schistosity but do not constitute the main orienting agent nor do they explain the development of new and characteristic minerals. Their effects become progressively less important as the reconstruction of the rock by crystallization proceeds. In some cases, however, they have been the main processes in forming fissile but not necessarily schistose rocks; for example, where limestones possess a good cleavage structure, it has arisen largely from rotation, enhanced by gliding.

Where a very high grade of metamorphism is reached, schistosity tends to be replaced by gneissic structure. This change is ascribed to two causes. First, the high temperature, while facilitating shearing movements, sets a constantly narrowing limit to possible shearing stress. Second, minerals of flaky or acicular habit belong for the most part to the lower and middle grades of metamorphism and are ultimately replaced by other minerals having no such distinctive habit.

MOVEMENT HYPOTHESIS

On this continent the chief advocates of the movement hypothesis are those who belong to what may be called the Sander school of thought, as many of their publications are based largely on Bruno Sander's *Gefügekunde der Gesteine*, which is regarded by them as the most authoritative treatise on the subject of structural petrology. The present writer has not read Sander's publications, having confidence that the English versions are fair presentations of Sander's ideas and methods—no doubt better than any the writer could make.

The emphasis which advocates of this hypothesis place on movement is shown by the following quotations. Knopf and Ingerson (1938, p. 26) say;

"Structural petrology is based therefore upon the idea that in the ordered arrangement of rock fabric we can see how the ultimate elements of the fabric have become oriented in response to the vectors of movement."

Knopf and Ingerson (1938, p. 42) state further that:

"The object of tectonic analysis is to correlate the rock fabric with different types of movements involved in the formation or deformation of rocks. . . . The most important identifying characteristic of movement is its symmetry, and this symmetry, which is clearly reflected in the fabric, is one way to recognize the fact that a certain kind of motion has given to the fabric its characteristic arrangement."

Of the various fabrics considered by structural petrologists, those of interest here are called tectonite fabrics, which are caused by rock flowage. Concerning the origin of tectonite fabrics, Knopf and Ingerson (1938, p. 179) write:

"Some tectonite fabrics . . . can be interpreted as the result of grain gliding, or of a combination of grain gliding and movement of rigid grains in a flowing medium. Others can be explained by a combination of rupture and of recrystallization from seeds whose position was determined by the orienting influences. Still other tectonite fabrics, mimetically crystallized, owe their preferred orientation to a post-tectonic crystallizing, which records the plan of the deformational movement, even intensifying it in some rocks."

The processes mentioned in this quotation may be subdivided into simple orienting mechanisms and processes involving mimetic crystallization, which is defined below. The simple orienting mechanisms are: twinning produced by deformation; rotation of rigid particles in a flowing medium, the particles being either the original grains or fragments produced by the rupture of the original grains; and gliding in certain directions along certain crystal planes, crystal lattices not suitably oriented for such gliding being brought into a suitable orientation by rotation or twinning. The processes involving mimetic crystallization, and the definition of this term, are explained in the following quotation from Knopf and Ingerson (1938, p. 39):

"Crystallization that reproduces any pre-existent anisotropy, bedding, schistosity, or other structures, is called *mimetic crystallization*. . . . A mimetic fabric is a fabric that records anisotropy either by recrystallization or by neomineralization, or by both. It can originate in two ways: (1) By the growth of nuclei that have acquired a preferred orientation during deforming and disintegrating movements, a process analogous to the annealing of cold-worked metals; (2) by crystal growth, where orientation is determined by the direction of easiest movement in the rock."

Knopf and Ingerson (1938, p. 67-68) explain the second type more fully as follows, taking a mica schist as an example:

"The biotite is concentrated in the plane of schistosity, which shows no lineation. The lack of tectonic orientation in the quartz grains that are associated with the biotite, and the generally fresh, undeformed nature of most of the biotite grains suggest that the micas did not assume their positions as a result of deforming

movements during or after their crystallization. The orientation in the mica is a mimetic elaboration of an already existent *s* plane. A fabric of this type commonly shows what has been called 'crystallization schistosity.' However, the crystallization in this fabric does not create the *s* planes by changing the shape of the pre-existent grains: it merely emphasizes an *s* plane that was already present."

The above quotations show that the Sander school attributes to movement all the oriented fabrics of deformed rocks. Thus flow cleavage would be included as a fabric expressing the movement involved in the deformation of the rocks containing it.

In the writer's opinion, however, the processes described in these quotations do not explain the development of typical flow cleavage. Considering the processes one by one, the critical arguments against attributing flow cleavage to them are as follows: (1) The simple orienting mechanisms, such as rotation, twinning, and gliding, do not explain the development of new minerals that is characteristic of rocks with flow cleavage. It may be remarked, however, that the analysis of these movements is of great value in understanding many of the features of cataclastic structures. (2) Crystal growth, occurring essentially as secondary enlargements of certain favored nuclei, whose arrangement is due to movement, is a concept that explains the oriented fabrics of some recrystallized rocks. However, this process, like the first, is unsatisfactory as an explanation of flow cleavage because it does not account for the development of characteristic new minerals. (3) The growth of micas and other new minerals along shear planes caused by movement would produce an oriented fabric only along certain planes. The oriented fabric would not be displayed by every plane parallel to the fissility, and thus the structure would not be that shown by typical flow cleavage. There is no doubt, of course, about this process being effective in some cases. For example it can produce certain features seen in modified types of fracture cleavage and it seems to afford an explanation of some gneissic structures which have cleavage parallel to the foliation.

At this point it is desirable to explain why the writer frequently uses the words, typical or type, in referring to some structures. The concept behind this usage is that a certain structure may reflect control by a certain process and that where only this process has occurred, or where it has been the predominating influence, the type structure is produced. In many cases, however, a rock displays some combination of the typical structures, which is only to be expected, because the processes forming the various deformative structures are closely related in origin. Thus, as a rule, typical flow cleavage is not the only structure displayed by a schist, but a study of many schists shows that they have certain features in common which serve to distinguish schistosity from allied structures, and these are the features which constitute typical flow cleavage and are

of critical significance in considering the origin of this structure. It follows that the more closely the structure of a rock approaches typical flow cleavage, the more valuable the occurrence of this rock is as an indication of the origin of this structure. This principle is of especial interest in the discussion of the field relations of flow cleavage which is given in the following section.

Of interest at this time is the fact that the conclusion reached above concerning the origin of flow cleavage does not mean that a rock possessing flow cleavage may not display certain textural features that are the result of mechanical movements. The writer, although not primarily concerned in this paper with structures other than flow cleavage, may add that he sees no reason why the statistical methods of fabric analysis developed by Sander cannot be used in conjunction with the methods advocated by Leith. The only necessity is that structures produced by orienting movements be distinguished from those produced by crystallization under differential stress.

From the viewpoint of the field geologist, the important point is the conclusion that flow cleavage can be interpreted as showing the stress condition in the rock when it crystallized. The writer believes that this principle is of great value and that it should not be discarded without very good reasons for doing so. Insofar as the writer can determine, such reasons have not been established.

RELATIONS OF FLOW CLEAVAGE IN FOLDED BEDS

RELATIONS ACCORDING TO RECRYSTALLIZATION HYPOTHESIS

Leith (1923) has been foremost in emphasizing the diagnostic value of the relations between bedding and flow cleavage in folded beds. The general principle is that flow cleavage is parallel to the axial plane of the fold containing it and is therefore inclined to the bedding. Quoting Leith (1923, p. 128) directly:

> "In wide observations through the pre-Cambrian rocks the writer has never yet found cleavage exactly parallel to the bedding for any considerable distance."

For comparative purposes it is worth noting that advocates of the recrystallization hypothesis believe that gneissic structure, on the other hand, may or may not be parallel to the bedding. Considering only gneisses of purely sedimentary origin, two types can be recognized. Regarding one type Leith (1923, p. 147) says:

> "The banding is due to original differences in the composition of the beds."

Harker (1932, p. 204) writes:

> "It is clearly suggested therefore that foliation is based ultimately upon pre-existent structures in the rocks. In a remote sense it is of the nature of a residual struc-

ture, though we should strain the meaning of that term by applying it where nothing remains of the original outlines. In the case of rocks of sedimentary origin, the prime cause of heterogeneity must be sought in stratification."

The second type of gneissic structure is regarded as a high-grade metamorphic equivalent of schistosity, the gneissic characteristics being due to segregation during recrystallization. In this connection Harker (1932, p. 203) writes that the bands:

"Show, at any one place, a common parallel orientation, which agrees with the direction of schistosity, if any, and is manifestly related to the same system of stress and strain in the rock. It is to be observed, however, that whereas schistosity declines as the highest grade of metamorphism is approached, foliation on the other hand often becomes more salient, and is developed on a larger scale."

The field observer who makes use of these principles would first determine whether a certain occurrence shows flow cleavage or gneissic structure. Then, if it is flow cleavage, he would base his interpretation on the principle that flow cleavage is parallel to the axial plane of the fold containing it. If it is gneissic structure, further study would be needed to determine what kind of gneissic structure is present. Parenthetically, the writer's experience is that gneissic structure in rocks of sedimentary origin is apt to be parallel to the original bedding, especially where the sediments have suffered a high grade of metamorphism without, however, being closely folded. But, even in such cases, foliation which crosses the bedding may be locally dominant. For example, in the Dinty lake area, north of Lake Athabasca, Saskatchewan, where the writer spent 2 months in 1938 mapping in considerable detail a small area, he concluded that the dominant foliation in the granitic gneisses reflects original bedding, although in a few places he noted foliation at an angle to the dominant type. In one of these places, both kinds of foliation were seen together. Here the less distinct banding outlines a sharp fold, and the more marked foliation is about parallel to the axial plane of this fold.

RELATIONS ACCORDING TO MOVEMENT HYPOTHESIS

The advocates of the movement hypothesis have developed certain generalizations concerning the relations between bedding and foliation, the latter term being used by them to include both flow cleavage and gneissic structure. These generalizations will now be discussed.

Fairbairn (1935) concludes that foliation developed parallel to the bedding may be distinguished from foliation parallel to the axial planes of the folds by a statistical study of the quartz grains in each case. In bedding foliation the quartz grains tend to have their *c*-axes parallel to one direction in the foliation plane. The interpretation is that gliding has occurred in the quartz grains parallel or subparallel to prism faces

and in the direction of the *c*-axes, and therefore the fabric indicates movement along the foliation planes. In axial plane foliation, on the other hand, the *c*-axes of the quartz grains tend to form a girdle, that is, they tend to be arranged like the spokes of a wire wheel, the axis of the wheel lying in the foliation plane. The interpretation is that the movement occurred along planes inclined to the foliation plane, the movement causing a dimensional arrangement of the grains in the rock and the foliation being due to this arrangement.

According to these ideas, movement did not occur along the foliation planes which are parallel to the axial planes of the folds. However, as noted below, other advocates of the movement hypothesis believe that foliation parallel to the axial plane is developed in slip folds, which involve movement along the foliation planes. Also, as far as the writer knows, all authorities agree that movement occurred along the parting planes in fracture cleavage, and these planes are usually parallel or subparallel to the axial planes of the folds. Thus there are reasons for believing that the methods described above for distinguishing between axial plane and bedding foliation would not apply to all cases.

In discussing the genetic relations between axial plane foliation and bedding foliation, Fairbairn (1935, p. 607) writes:

> "As pure hypothesis, it is held by a number of workers in Europe that axial-plane foliation may arise as an early stage of deformation of sediments, and that bedding foliation is a later development. Although no proved case of a transition between the two types is known to the writer, the conception is not without value. It is entirely possible, for instance, that an early-formed axial-plane foliation should become the locus of later shearing movements, and that with continuing deformation the bedding should approach parallelism with the shearing direction by a combination of slicing and rotation. At an advanced stage all trace of the earlier folded structure is obliterated and the remaining primary structures, bedding, etc., lie parallel to the former axial-plane foliation, now shear surfaces, and here called bedding foliation."

As the writer understands it, this quotation calls on continuing deformation, which involves a progressive change from open to closed folds, as a means of bringing the bedding into parallelism with the foliation. It is true, of course, that if one considers the exposures individually parallelism between bedding and foliation becomes the rule in most outcrops as the folding approaches the isoclinal type, the more so because the axial planes of the folds are apt to be poorly exposed, and the bedding in the vicinity of the axial planes is likely to be indistinct. There is thus a good chance of finding, as Fairbairn says in the above quotation, that: "The *remaining* primary structures, bedding, etc., lie parallel to the former axial-plane foliation"; (the italics are the present writer's). But the writer fails to see how continuing deformation can change the relations between the foliation and the axial planes of the folds, when these relations are considered collectively and on a sufficiently large scale.

From the viewpoint of the geologist who is mapping the regional structure of an area, the bedding of closed folds is no more parallel to axial plane foliation than is the bedding of open folds. Therefore the writer concludes that the type of so-called bedding foliation which is produced by the method described in the foregoing quotation is actually not bedding foliation at all.

This discussion emphasizes a difference in viewpoint which may be important in explaining certain disagreements. The writer has often wondered whether some geologists who describe cases where bedding and cleavage are parallel are referring to the relations seen in most or all outcrops or to those which apply to the true structure of an area of sufficient size to contain one or more complete folds. It is quite clear, of course, that Leith (1923) and most structural geologists refer to the relations that are disclosed when bedding and cleavage are traced, either directly or by circumstantial evidence, over a considerable distance, at least far enough to include a large part of any fold that may be present. The writer's own experience has led him to be very suspicious of interpreting as bedding any foliation which maintains a constant attitude over a large area, regardless of the relations that may be shown by individual outcrops.

Knopf and Ingerson (1938, p. 157-161) describe two types of folds. Slip folds are produced by displacements of progressively increasing or decreasing amounts along a single set of slip planes which are inclined to the bedding. To the writer, the process seems similar to that which produces drag near some faults. Flexural-slip folds occur where there is an alternation of competent and incompetent layers, flexure taking place in the competent layers, while slipping movements develop in the intervening incompetent layers as they are bent. It is apparently agreed that flexural-slip folding is the kind that occurs most commonly in bedded rocks.

Regarding the oriented fabrics in these two types of folds, these authors say that in slip folds the platy minerals, such as mica, are oriented in the slip plane transverse to the curve of the fold. An illustration they give shows an arrangement of mica in slip planes almost parallel to the axial plane of the fold. In flexural-slip folds the platy minerals follow the arch of the fold. In short, rock cleavage produced by the parallel orientation of platy minerals is about parallel to the axial plane in slip folds and parallel to the bedding in flexural-slip folds.

These authors further note the following differences between slip folds and flexural-slip folds. In slip folds, the thickness of a bed, measured along the slip planes, is constant. As a result, the geometric thickness of all beds is less on the limbs of a fold than on its crest or trough. In

flexural-slip folds, the competent beds tend to maintain their true thickness, while the incompetent beds tend to be thinner on the limbs of a fold than on its crest or trough. In slip folds the slipping movements occur in all the beds, whereas in flexural-slip folds they are confined to the incompetent beds. In slip folds the folding shows no relation to the competency of the beds, whereas in flexural-slip folds the incompetent beds develop smaller folds than the competent beds.

The writer's comment on these statements is that the attitude of rock cleavage determined by the parallel orientation of platy minerals is not always in accord with the relations described. The writer, and doubtless every other field observer, has seen folds which, although having all the characteristics of the flexural-slip type, show cleavage that is about parallel to the axial planes of the folds. Typically, this cleavage is confined to the incompetent beds, which alone makes it impossible to classify the folds as slip folds. It is therefore concluded that field observations do not support the generalization that all folds which have cleavage parallel to their axial planes are slip folds.

FIELD RELATIONS AS AN INDICATION OF ORIGIN OF FLOW CLEAVAGE

The origin of flow cleavage has already been discussed in this paper, the conclusion being that flow cleavage is due to recrystallization under differential stress and not to mechanical movements during deformation. In arriving at this conclusion, no restrictions were placed on the amounts or the spatial relations of the movements. Now, in some simple folds it is possible to estimate, roughly at least, the amount of deformation that occurred in certain beds having flow cleavage. Such occurrences enable one to apply approximately quantitative tests to the movement hypothesis and the recrystallization hypothesis. In the following paragraphs the writer will attempt to show that an analysis of this sort gives further support to the conclusion already reached concerning the origin of flow cleavage.

In well-exposed areas where originally sandy and shaly beds have been folded at moderate depths, complete folds can occasionally be seen, and the existence of the following relations can be established: (1) The flow cleavage is virtually confined to the slaty beds, which often show minor drag folds; (2) it is parallel to the axial planes of the folds; and (3) it does not change noticeably as it is traced across the crest of a fold, nor is its development appreciably better on closed folds than on open folds.

Of these relations, the first one shows that the folds are of the common or so-called flexural-slip type, that is, the deformation of the slaty beds is due largely to forces transmitted to them by the flexing competent

beds, which tend to maintain their original dimensions. Thus a competent bed tends to be flexed into a simple fold, with no extension or shortening (of its neutral plane in the ideal case) parallel to the axis of the fold or in any direction along the bedding. Flexing in this manner, the competent beds slide over one another, the upper bed in an anticline moving directly toward the axis of the fold with respect to the lower bed. The difference in curvature between the top and bottom surfaces of the competent beds creates potential spaces on the crest of the fold, such spaces being occupied by incompetent material that has flowed in from the limbs. The entire deformation produced in an intervening slaty bed can be visualized from a cross section of the fold.

The reader may object because this concept neglects some complicating factors that may be present, such as movements parallel to the fold axis, lateral migration of the axial plane during folding, forces due to the weight of overlying material, and forces due to shoving actions by the flexing competent beds. The answer is that, to be of significance in considering the occurrence of typical flow cleavage, any such complicating factor must have general application. Factors such as those just mentioned do not meet this requirement. It is true that the forces due to gravity and shoving actions are generally present, but their net effectiveness will vary in different cases, depending on the depth of burial, the thickness and strength of the competent beds, and so forth. For example, a slaty bed need not be subject, in all cases, to the weight of the overlying beds during folding.

Accordingly we may accept, as critical evidence, the deformation shown by a cross section of a simple fold. Considering first the movement pattern, it is evident that, regardless of how the movement planes are oriented with respect to the fold, the amount of movement must be greater on the limbs than on the crest of a fold, it must be greater at any one place on a fold if the fold is closed than if it is open, and it must vary with the relative thicknesses of the competent and the incompetent beds that are involved. Insofar as the writer is aware, flow cleavage structures do not display systematic differences that are in accord with the above analysis of the movement pattern. Considering next the absolute amount of movement, we may look at the crest of a simple fold for critical data. Here slipping between beds is at a minimum. The direction of greatest elongation is perpendicular to the bedding, and its amount is shown by the thickening of the incompetent bed. To the writer it does not seem possible that in some simple folds, this amount of strain could cause rotation, twinning, and gliding sufficient to develop the high degree of mineral orientation that is characteristic of flow cleavage in slaty beds.

At first thought it might appear that the same objections would apply to the concept that fracture cleavage is produced by mechanical movements. However, the movements producing fracture cleavage can be concentrated along planes that are far apart in comparison to the grain size of a slate, and the amount of movement needed to produce a fracture need be only very small. Moreover, fracture cleavage often shows better development as the degree of folding increases, which is in accord with the requirements of the movement hypothesis.

Parenthetically, it may be remarked that, to the extent that fracture cleavage grades into flow cleavage, the structures in intermediate types may be partly or largely due to orienting movements. This consideration, however, does not detract from the conclusion that typical flow cleavage is not due to orienting movements.

Turning now to the recrystallization hypothesis, the writer can see no equivalent difficulties in using this hypothesis to explain the relations between bedding and flow cleavage in typical occurrences. There is no reason for believing that the oriented growth of minerals under the influence of differential stress requires a great amount of distortion or even a great amount of differential stress. Many examples are known where an oriented crystal growth has taken place under the influence of what were apparently only minor stress differences. Also, across the crest of a simple fold the stress condition is essentially uniform, consisting largely of compression normal to the axial plane, together with more or less hydrostatic pressure. Such a stress condition would account for the uniform character and attitude of flow cleavage in the vicinity of the crests of folds.

It is perhaps worth while to call attention to the fact that these conclusions are based largely on observations made near the crests or troughs of folds. On the limbs, especially those of closed folds, the amount of shearing movement may be very large, and the relations there have not the same diagnostic value. For the field geologist, the relations near the crests or troughs of folds are the ones that are most useful, particularly if the folding tends to be isoclinal. The relations seen on the limbs are as a rule of value only as indications of what to expect and where to look for critical information. Thus, from the viewpoint of the field geologist, the failure of the movement hypothesis to explain typical relations on the crest of a fold is a serious defect.

RELATIONS IN COMPLEX FOLDED STRUCTURES

Where the folding is simple, field observations have established the validity of the general principle that flow cleavage is parallel to the axial planes of the folds. Where the structure is complex, however, apparent

contradictions of this principle may be encountered. The writer believes that most of these apparent contradictions are due to lack of complete knowledge of the field relations or to the mere weight of natural complications rather than to any failure of the general principle. To support this viewpoint, as well as to help those who encountered problems similar to some that the writer has met, this paper will be concluded by describing a few common situations where the relations are apt to be misleading.

First of all, it is worth while noting that the effect of variations in the competence of the beds is an important consideration. Places where the incompetent beds form relatively thin layers between the competent strata are the most favorable for observation because the deformation of the weaker beds is subject to a broad control imposed by the stronger ones. On the other hand, thick zones composed dominantly of incompetent material, with a few thin competent layers, are unfavorable because faulting and folding of the competent layers produce a great variety of local stresses. Within such zones, the minor folds and cleavage structure may have no direct relation to the major structure. Finally, where the whole succession is composed dominantly of incompetent beds, there may be no general pattern to the structure, and any deductions which assume the presence of such a pattern will be liable to serious error.

Turning now to rather special situations, the writer will describe a few examples which are of interest in this connection. Several of the examples are taken from the Yellowknife district, N. W. T., not because the structures there are considered unique but because the unusually complete exposures often permit one to observe every detail of the folded beds.

One type of situation involves a warping of the whole structure, including the bedding and the pre-existent cleavage. A good example was seen by the writer near the Fraser vein of the Thompson-Lundmark property in the Yellowknife district, N. W. T. Here the bedding has a general north-northwest strike and dips about 45° easterly. The relations between bedding and flow cleavage indicate the presence of folds having one limb with the attitude given above and the other limb striking nearly due east and dipping very steeply. In studying the area, however, the first fold the writer saw was of a type that does not fit this pattern. It could be described as a monoclinal flexure, the strike of the beds forming a reverse curve by swinging from north-northwest to west-northwest and then back again. Finding this fold was unexpected, but closer examination showed that the cleavage suffered an equivalent change in attitude, indicating that both bedding and cleavage had been warped. On being traced into higher beds, the flexure gradually changed into a small cross fault. Obviously this structure represents merely a minor and late

period of deformation. The cleavage and bedding relations in its vicinity cannot be used, in their present attitudes, as a means of determining the character of the major folds in the area. It may be added that further study revealed the presence of a larger fold, of the type that would be expected from the normal cleavage and bedding relations in the area.

Another situation involves faulting or shearing along definite zones, with the development of drag folds and cleavage which reflect the movement in these zones, and not the major folded structures of the region. Minor movements of this sort can easily be discounted in exposures of ordinary size, but the effects of larger movements may be difficult to recognize, at least in the early stages of mapping. The following paragraphs describe two examples that fall in this class.

One example involves thrusting along the short limb of an asymmetric fold, in a direction opposite to that of the slip between beds that is normally dominant. Naturally the drag folds and cleavage relations on this limb will be abnormal. Where these conditions are present it is quite possible, when interpretations are based entirely on the relations between cleavage and bedding and on the shapes of the drag folds, to interpret as a homoclinal succession a structure that is really close to the imbricate type and involves much duplication of strata.

The second example, which involves shearing oblique to the axial plane of a major fold, is given by Ambrose and Gunning (1940, p. 40-45). The structural relations they describe can be summarized as follows. The major structure is a very gently pitching syncline whose limbs dip steeply and are virtually parallel to one another. The presence of this major structure was inferred from determinations of the stratigraphic tops of the sedimentary and volcanic formations in the area. The drag folds and associated fracture cleavage do not show the relations to the syncline that would be expected if they were developed as part of the synclinal structure. For one thing, the pitch of the drag folds is steep, in most cases nearly vertical. Thus the axes of the drag folds are about perpendicular to the axis of the major syncline. Furthermore, drag folds showing the same direction of movement occur on both limbs of the major syncline, which would be impossible if they were developed as part of the major structure. The conclusion is that the drag folds and associated cleavage were produced by horizontal movements later than, and unrelated to, the isoclinal folding that caused the major synclinal structure. Therefore the relations of cleavage to bedding, and the drag folding, cannot be used to determine the character of the major folded structure in this area.

A third situation also involves two periods of deformation, but in this case the effects of neither deformation are confined to a definite zone.

This situation is illustrated in several parts of the Yellowknife district, N. W. T., where the writer observed a generally flat cleavage as well as a generally vertical one. For example, on the Try Me group of claims, the dominant structure is a tight anticline. One limb strikes northwest and is nearly vertical. The other limb strikes about due north and dips 45° E. In many places, especially on the vertical limb, flow cleavage is well developed in the slaty beds and is about parallel to the axial plane of the fold described above. In fact, the character of the fold was first inferred from the bedding and cleavage relations seen on the vertical limb, this inference being confirmed later by observations on the crest of the fold. However, there is also a nearly flat cleavage, parallel to the axial planes of minor folds which are most noticeable on the gently dipping limb, although also seen on the other limb. In cases of this sort, a complete explanation of the structure requires detailed areal mapping, and, in advance of this, interpretations based on local observations may be misleading. The writer has the impression that the flat cleavage on the Try Me group reflects an early deformation that was widespread but not necessarily intense, but this concept must be classed as almost purely hypothetical.

Finally, a situation can be described where the abnormal relations must be ascribed to local failure of the general principle rather than to the presence of two distinct periods of deformation. An example was seen on the Pan group, also in the Yellowknife district, N. W. T. Here the dominant structure consists of large drag folds that pitch steeply northwest. The prevalent cleavage has normal relations to these folds. However, in places there are thin layers with abnormal cleavage. Such layers may alternate with other layers that show normal cleavage relations with respect to the major folds. The abnormal cleavage is nearly perpendicular to the axial planes of the major folds, although, like the normal cleavage, intersecting the bedding along lines parallel to the axes of the folds. Also, there are minor drag folds whose axial planes are parallel to the abnormal cleavage. They indicate movement betwen beds in a direction that is exactly opposite to that which is usually present in folding. These abnormal relations are only minor features, which would not confuse anyone who based his conclusions on the prevalent structures. Their origin is not clear, but it seems that they were caused by conditions which locally reversed the direction of slip between beds. Possibly, in the late stages of folding, the crests or troughs formed by certain beds were pushed, like a piston, into the crests or troughs of adjacent beds, the movement being due to the collapse of competent arches which had hitherto tended to create spaces between these crests or troughs.

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DEFORMATION LAMELLAE IN QUARTZ FROM THE AJIBIK FORMATION, MICHIGAN

BY H. W. FAIRBAIRN

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ABSTRACT

Petrofabric study of a specimen of quartzite from a shear zone in the Ajibik formation of the Marquette district, Michigan, contributes the following to the quartz lamellae problem:

(1) Excellent orientation relations of the deformation lamellae and crystal axes indicate the control of the quartz orientation by the lamellae rather than by the axes. A translation hypothesis is considered to explain the relations best.

(2) A fixed glide line $[m : r]$ and variable glide planes containing $[m : r]$ satisfy best the irrational lamellae planes and the restricted orientation of the axes associated with them.

(3) The character of the lamellae orientation leads to the tentative conclusion that the strain was nonrotational.

(4) The possibility is mentioned that some of the quartz maxima in recrystallized tectonites may be remnants of axis concentrations related to lamellae which have since been obliterated by the recrystallization.

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INTRODUCTION

The deformation lamellae of quartz, long known to European petrologists, have always excited considerable curiosity and have given rise to speculations, more or less sound, as to their origin. Although they are no longer the mineralogic mystery they once were, the problems which they offer to petrofabric investigators are still a challenge. A surprising number of petrographers have never seen them or, having seen them, have failed to note their invariable habitat in nonrecrystallized, highly deformed rocks. They are not to be confused with twin lamellae, well known in quartz, but not resembling deformation lamellae in any way. There is no evidence that twin lamellae are due to deformation. Usually they are not visible unless the grains are etched.

Deformation lamellae are commonly referred to, erroneously in most cases, as "Boehm" lamellae. The original description by Boehm (Griggs and Bell, 1938) refers, however, to lines of inclusions in the quartz of recrystallized rocks which may or may not be relicts of lamellae. Due to this ambiguity in nomenclature, the prefix "Boehm" should be discarded.

As it is not easy to obtain material suitable for a petrofabric analysis, most references to deformation lamellae are brief and the interpretations inconclusive. In the present study a specimen of Ajibik quartzite from the Marquette district was used, and the statistical analysis led to conclusions which, though incomplete, warrant presentation at this time.

FIELD RELATIONS AND PETROGRAPHY

The specimen analyzed here by petrofabric methods comes from a shear zone a few feet wide and of undetermined length. Figure 1 of Plate 1 shows a typical specimen of the greenish, micaceous quartzite. A pronounced foliation is shown by parallel mica foils and elongate quartz eyes. The mica-rich layers interspersed with quartz-rich layers have a pronounced lineation parallel to the foliation. One of these is shown at the base of the specimen in Figure 1 of Plate 1. This lineation is confined to the shear zone and is perpendicular to the regional lineation usually denoted in petrofabric analysis by the letter *b*. The maximum dimensions of the elongate grains are not shown in the photograph since its plane is perpendicular to the lineation.

FIGURE 1. SHEARED QUARTZITE FROM THE AJIBIK FORMATION, MARQUETTE, MICHIGAN
About three fourths natural size. Note prominent quartz eyes. Their maximum dimension is perpendicular to the plane of photograph. Two quartz stringers show as irregular dark lines. A mica-rich layer forms base of specimen.

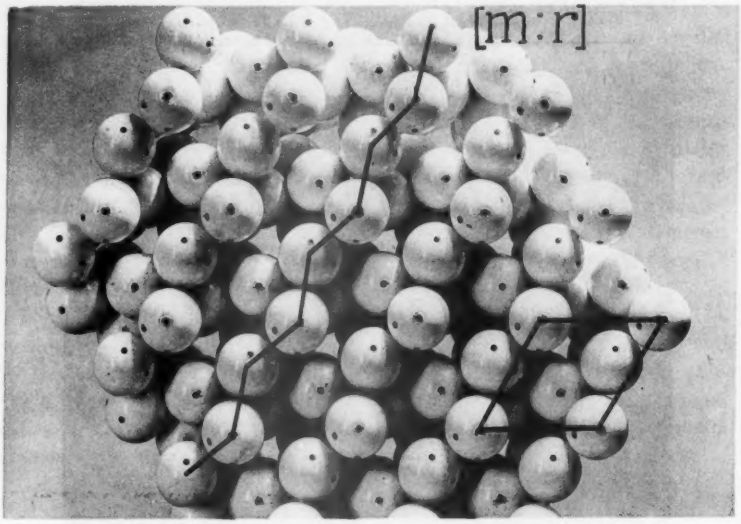

FIGURE 2. PACKING MODEL OF LOW QUARTZ
Looking parallel to crystal axis. Parallelogram shows dimensions of a unit cell. Zigzag line connecting successive *O* atoms marks the direction [m:r]. Model constructed by Butler and Buerger.

AJIBIK QUARTZITE AND PACKING MODEL OF LOW QUARTZ

FIGURE 1. WELL-DEVELOPED LAMELLAE IN A LARGE QUARTZ GRAIN
s denotes foliation in thin section. Crossed nicols. x 140.

FIGURE 2. COMBINATION OF LAMELLAE AND UNDULATORY ZONES
Note angular relation between them. Foliation as in Figure 1 above. Crossed nicols. x 140.

PHOTOMICROGRAPHS OF QUARTZ LAMELLAE

In thin section the impression gained from preliminary study of the specimen is heightened. Quartz eyes vary from the centimeter sizes of Figure 1 of Plate 1 down to .05 mm. The elongate character persists in general but is slightly less pronounced in the smaller grains. Grains averaging about 0.5 mm. are perhaps more abundant than other sizes, and on them the main part of the petrofabric analysis is based. Most of the quartz eyes appear as islands, each isolated from its neighbor by an interwoven network of mica shreds. The long dimension of these mica foils is usually less than that of the quartz islands which they enclose. They give the effect of a beard to the quartz eyes, either terminating abruptly against or projecting into the grains.

Almost without exception, the quartz shows undulatory extinction. In grains large enough to measure successfully, this extinction shows parallel zoning whose trend is subparallel to the crystal axes. Less conspicuous are the lamellae which occur in a considerable proportion of the quartz eyes. Figure 1 of Plate 2 shows a quartz grain in which they are particularly well developed. They consist of narrow, bright lines which resemble a series of welts raised on the surface of the grain. The Becke line test is not always conclusive, but many of the lamellae appear to have a lower index than that of quartz. They are statistically parallel in any one grain but in detail show small deviations from parallelism. Their width varies slightly, and they taper at the ends. They may extend entirely across the grain as in Figure 1 of Plate 2 or may terminate without making any contact with the grain boundary. Lamellae occur in aggregates rather than singly. They show also an important relation to the undulatory zones with which they are usually associated (Pl. 2, fig. 1). In most cases the two structures are seen to intersect at approximately 90 degrees. This and other features of quartz lamellae have been previously described by Sander (1930).

Neither rock nor thin sections show any sign of recrystallization. The variable grain size of the quartz, its irregular, jagged boundaries, and the presence of undulatory zones and associated lamellae provide adequate evidence of this. The minute shreds and felted aggregate of mica flakes scattered in and around the quartz islands show likewise the absence of recrystallization processes.

PETROFABRIC ANALYSIS

Measurement of deformation lamellae of this type cannot compare in accuracy with that of normal twin lamellae or cleavage surfaces but gives nevertheless, if doubtful measurements are discarded, a good approximation. In many cases the lamellae are found only upon suitable manipulation of the universal stage and disappear completely from view

with a few degrees' rotation. This accounts in part for the scant reference made to them in ordinary petrographic work done without the universal stage. Slight deviations from parallelism with the line of sight of the observer render them invisible. Too much illumination also makes them

FIGURE 1.—*Collective diagram of 337 lamellae poles*

(*s*) foliation; (*a*) lineation; (*L*) two statistical lamellae positions. Contours 8-6-4-2-1-0%.

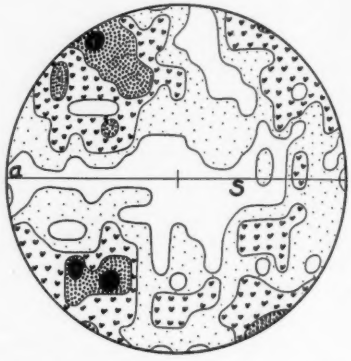

FIGURE 2.—*Collective diagram of 364 axes*

Includes only those grains having lamellae. Number of lamellae poles (337) in Figure 1 does not correspond with number of axes because some of former were too difficult to measure even approximately. *a* as in Figure 1. Contours 4-3-2-1-0%.

difficult to see, and it is usually advisable to cut down the field of view for best results.

The diagrams shown here were obtained from measurements made on two sections normal to each other and both perpendicular to the foliation. (*See* Figure 4.) One is parallel, the other is normal, to the lineation. The results are combined into one set of diagrams with the lineation *a* represented by the east-west diameter.

Figure 1 is a collective diagram of lamellae poles which shows two prominent maxima about equidistant from the foliation surface *s*. The average lamellae positions corresponding to these pole maxima are designated by *L*. Figure 2 shows the orientation of the axes associated with the grains of Figure 1. Maxima appear in the same quadrants but are much less pronounced than those of the lamellae poles. Figure 3 shows the orientation of all the quartz axes measured in the sections, with and without associated lamellae; it shows much more spreading of the axes than Figure 2.

The angular relations of individual lamellae poles and quartz axes are compiled in Figure 7. As indicated by Figures 1 and 2, quartz lamellae do not form parallel to a rational crystallographic plane. Figure 7 shows that the orientation is not entirely random, however. About

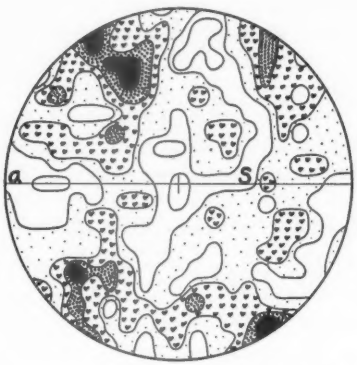

FIGURE 3.—*Collective diagram of 613 axes*

No selection of grains was made. *s* and *a* as in Figure 1. Contours 2.5-2-1.5-1-.5-0%.

80 per cent of the lamellae poles lie between the 7- and 36-degree limits. These facts have long been known, and the present data merely substantiate them more fully.

There is a further significant relation between the axes and lamellae poles. In Figure 2 the axes are not randomly oriented in a wide zone encircling the lamellae pole maxima. Instead, the two concentrations lie approximately in a statistical crystallographic zone with respect to the lamellae poles. This zone has an axis parallel to the lamellae and approximately normal to the crystal axis. The relation is pictured schematically in Figure 5. For either of the two ideal orientations shown (full line or broken line) the statistical crystal must fulfill two conditions in order to account for the existing orientation. (1) It must have restricted freedom of orientation about one particular horizontal [1] axis. The degree of restriction is indicated by the varying angles in Figure 7. This horizontal axis is normal to the foliation and parallel to the lamellae. For reasons to be presented later, the crystallographic

[1] "Horizontal" refers here to crystallographic axes and is not to be confused with fabric or geographic axes.

line [m:r] [2] has been selected as the most likely horizontal axis. (2) It must have no freedom of orientation with respect to any other horizontal axis. If it had, the axes would encircle the lamellae pole maxima in a broad zone.

Figure 6 shows the extent to which this ideal condition is fulfilled. By construction from Figures 1 and 2 the orientation of those lines in the lamellae which are also perpendicular to the crystal axes was obtained. The points of emergence of these lines are concentrated parallel to the average lamellae positions L, as would be expected. In addition, most of the chief maxima lie in the neighborhood of the perimeter of the diagram. There is thus a reasonably good correspondence between the ideal relation assumed in Figure 5 and the actual condition shown in Figure 6.

DISCUSSION OF ORIENTATION

Comparison of Figures 1, 2, and 3 shows that the lamellae orientation is much more critical than the axes orientation. Lamellae developed only in grains whose axes are oriented in a particular way with respect to the direction movement. (Compare Figures 2 and 3.) In grains not possessing this favorable orientation no lamellae developed. As this was the case for about one-half the grains measured, the inadequacy of an analysis of these highly deformed rocks based on axes measurements alone is clearly indicated.

Since the lamellae occur only in sheared rocks it is reasonable to assume their development to be due to deformation. This must remain a geological assumption as no experimental lamellae have been produced. One must therefore consider in turn the probability of deformation by twinning, rupture, and translation. With regard to the first, the irrational crystallographic orientation of the lamellae and their optical continuity under crossed nicols make a twinning hypothesis untenable.

With regard to the second, the lack of actual ruptures associated with the lamellae is remarkable. If it is assumed that the lamellae are healed shear fractures, then the curved character of some, their lenslike shape, and termination in many cases inside the grain boundaries do not characterize this type of rupture. The close spacing of the lamellae and orientation with respect to the main s-surface excludes their origin as tension ruptures.

The remainder of the discussion is concerned with the translation hypothesis. Although numerous attempts have been made, particularly by Griggs, no translation has been produced artificially in quartz. This

[2] Crystallographic directions or lines are referred to by combining the letters for two planes whose line of intersection represents this direction. Thus [m:r] is the direction of the intersection line of the prism m and the rhombohedron r.

final proof, although desirable, need not discourage hypotheses which are dependent on field and petrofabric evidence alone. The evidence assembled in the present study offers better support for translation than any presented heretofore. Quartz, as already pointed out (Fair-

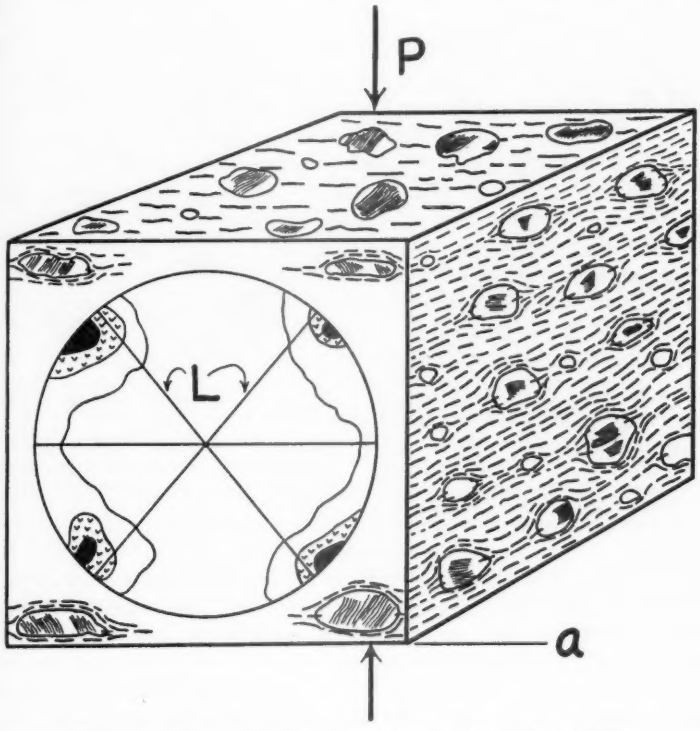

FIGURE 4.—*Block diagram to correlate petrofabric and field data*

(*a*) lineation; (*L*) average lamellae positions in the quartz; (*P*) probable orientation of chief stress and direction of shortening. Top surface of block is foliation surface. Parallel lines in quartz eyes represent deformation lamellae.

bairn, 1939a), is not an ordinary mineral from the standpoint of deformation. Extraordinary difficulties confront the analyst who would rationalize its experimental behavior in terms of its crystal structure. Following the methods of analysis successful for other minerals, quartz should have no glide lines or glide planes. The high degree of interlocking of the *Si-O* tetrahedra renders it impossible to deform quartz beyond the elastic limit without disrupting at least a few of the very strong *Si-O*

bonds. Deformation planes develop in related *Si-O* tetrahedral type minerals only because these planes avoid the *Si-O* bonds. It was suggested, however (Fairbairn, 1939a), that the direction $[m{:}r]$ in quartz was the most likely to be utilized if translation did occur. This is

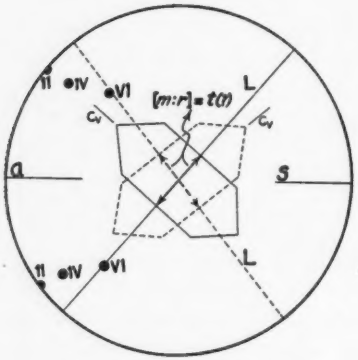

FIGURE 5.—*Diagram showing how translation parallel to* $[m{:}r]$ *may satisfy orientation relations*

Full lines show one orientation of an ideal crystal; broken lines show the other. c_v is vertical crystal axis in each case. $[m{:}r] = t$ (?) is assumed glide direction in each crystal.

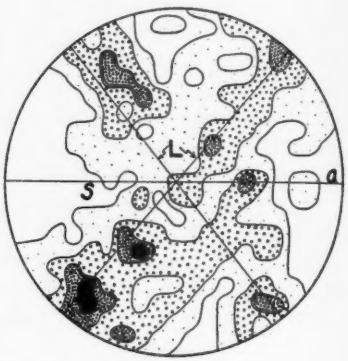

FIGURE 6.—*Collective diagram of 337 lines*

(Assumed to be $[m{:}r]$) which are parallel to lamellae and perpendicular to axes. *a*, *s*, and *L* as in previous figures. Contours 4-3-2-1-0%.

illustrated in Figure 2 of Plate 1 where the path followed by gliding *O* atoms is shown by a zigzag line. No attempt was made to assign a glide plane to this $[m{:}r]$ direction. This analysis of the crystal structure may have importance in the present example. If the significant axis normal to the crystal axis in Figure 5 be assumed to be $[m{:}r]$, a rational explanation of the lamellae-axes relations is possible in terms of a translation mechanism. Quartz grains having $[m{:}r]$ directions as in Figure 5 ($[m{:}r]$ occupies three directions in each ideal crystal) may glide parallel to this direction and utilize as a glide plane any surface in the crystal subparallel to the base, preferably in the range 7 to 36 degrees (Fig. 7). This may possibly account for the irrational relation of the lamellae to the axes as well as their high degree of orientation. Figure 6, showing the orientation of lines in the lamellae which are perpendicular to the axes of Figure 2, is considered a diagram of $[m{:}r]$ lines. There is no proof that these are $[m{:}r]$ rather than any other lines fulfilling the same condition, since the optical methods used for thin sections do not give a unique orientation for uniaxial crystals.

Precedent for the assumption that a fixed glide line and variable glide lines can function as a translation mechanism is given by Taylor and Elam (1935) in experiments carried out on iron and β-brass crystals. They found an invariable glide line parallel to [111] associated with glide planes of variable orientation, the latter being controlled only by the orientation of the stress. This evidence is not presented here because of any close analogy to the deformation of quartz but merely to indicate that there is experimental evidence which deviates from the traditional fixed translation plane idea which dominates most interpretations. The evidence of the quartz lamellae described in this paper is the first to be presented for a rock-forming mineral which leads to the fixed glide-line, variable glide-plane conception of deformation.

The boundaries of the quartz grains show no direct evidence of translation where the lamellae make contact with the edges of the crystals. The grains are completely enveloped in mica shreds, however, many of these projecting into the quartz eyes. Steplike offsets along the grain boundaries due to translation would be exceedingly small in any case, and the masking effect of the mica rendered the search unsuccessful. On the other hand, the elongate character of most of the quartz and its dimensional orientation parallel to *a* provide suggestive thin section evidence of an elongating process by translation along the inclined lamellae.

Mention should be made at this point of the possible evidence of translation to be derived from the relatively low refractive indices of the lamellae. In theory at least, plastic deformation decreases density and should therefore decrease refractive indices. The changes are believed to be exceedingly minute and difficult to detect. Becke (1892) stated that the lamellae refraction was less than that of quartz. There is some evidence of it in the Ajibik lamellae, but the writer does not consider it conclusive.

The discussion thus far has not included the undulatory extinction bands perpendicular to the lamellae. It does not seem possible to state definitely whether these developed before, contemporaneously with, or later than the lamellae. They represent crystal distortion and are known to be associated with ruptures trending parallel to the extinction zones, *i. e.*, subparallel to the crystal axes (Griggs and Bell, 1938). Where a bend occurs in the lamellae the extinction bands passing through them diverge so as to maintain the perpendicular relation. It seems reasonable to suppose that the two deformation phenomena developed simultaneously, one definitely by elastic distortion and rupture, the other presumably by translation. In Figure 5, therefore, translation would occur parallel to the two [*m*:*r*] directions and rupture subparallel to the

two c_v directions, each being an expression of the easiest type of deformation possible in the given crystal direction.

Insofar as the writer knows, quartz is unique in its possession of deformation lamellae having the restricted crystallographic irrationality shown in Figure 7. Beyond any doubt this property is explainable in terms of its crystal structure but will not be attempted here. On the

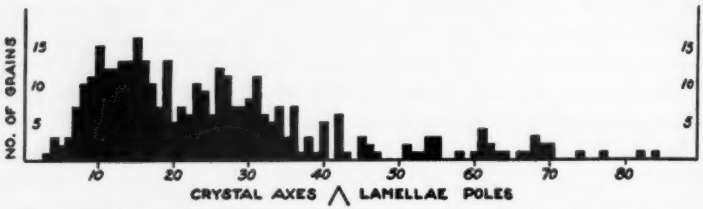

FIGURE 7.—*Graph showing angular relations of axes and lamellae poles*
Accumulation of measurements between 7 and 36 degrees is conspicuous.

other hand, undulatory extinction due to deformation is not uncommon in other minerals such as calcite, feldspar, and mica. The hypothesis of translation favored here for the lamellae does not close discussion of the problem by any means; there are undoubtedly other facts awaiting discovery if the right material is found.

RELATION TO SHEAR ZONE

The data just presented are confined to a single specimen and cannot therefore be extrapolated to interpret the entire structure of the shear zone. The chief object of the analysis, moreover, was to study quartz lamellae and their origin. However, certain features of the orientation are significant in their relation to the field structure. Figure 4 shows these schematically. L in the inset circle represents the two average positions of the lamellae with respect to the foliation (top surface of block) and the lineation a. If the quartz lamellae developed by translation, as postulated here, then the elongate quartz eyes and parallel mica foils forming the lineation represent also the elongation direction in the shear zone. The direction of shortening would thus be perpendicular to the foliation (shown by arrow). The equal distribution of the lamellae in the two maxima of Figure 4 and the approximately equal angles (39 and 41 degrees) which they subtend with the assumed direction of shortening indicate that the chief stress P was applied in a direction parallel to the bisectrix of the lamellae, or parallel to the shortening. If P were applied obliquely, unequal lamellae development in the two

maxima and asymmetry toward the foliation would be expected, that is, nonrotational rather than rotational strain is indicated.

Since the lineation in the shear zone studied here is perpendicular to the lineation usually developed parallel to fold axes on a regional scale, there might be further discussion of the general problem of elongation, first outlined by Sander (1930) and later by Fairbairn (1936). However, in view of the limited geological character of the present study, this phase of the deformation cannot be profitably discussed at this time.

RELATION TO RECRYSTALLIZED ROCKS

In two recent papers (Griggs and Bell, 1938; Fairbairn 1939b) a fracture hypothesis for quartz axes orientation has been developed. Since it dealt almost entirely with recrystallized rocks the matter of lamellae orientation was not considered. Three of the maxima discussed under the fracture hypothesis fall, however, within the range of the axes maxima associated with the lamellae in the present investigation. These are shown as II, IV, and VI in Figure 5. They have been omitted from the right half of the diagram. The significant axes-lamellae relation discussed in the preceding paragraphs immediately raises the question as to whether these particular axes orientations of recrystallized rocks are not relict orientations of previous deformations in which lamellae were produced. Recrystallization obliterates visible evidence of deformation such as lamellae and undulatory extinction zones but does not appreciably disturb the crystal orientation insofar as we know. Thus concentrations of axes in the neighborhood of the ideal maxima II, IV, and VI (based on a fracture hypothesis) may represent nothing more than ghosts of an earlier deformation in which lamellae parallel to an $[m{:}r]$ direction functioned as a translation system.

PREVIOUS INVESTIGATIONS OF LAMELLAE

Although numerous studies have been made of quartz lamellae only one statistical analysis has been published up to the present. Sander (1930) presented a diagram (D. 21) of quartz lamellae and of the associated axes (D. 19) which, however, does not suggest the significant relations found in the Ajibik quartzite. He interpreted the lamellae as translation surfaces. Becke (1892) concluded at first that they represented incompletely healed fractures but later he agreed with Sander. Mügge (1896) was the first to note the divergence of the undulatory extinction zones where they cross bends in the lamellae, thus maintaining the perpendicular relation at all points. He believed that the lamellae were due to axial translation (bend-gliding) and that the undulatory

extinction represented contemporaneous distortion of the crystal. He presented no petrofabric analysis or crystal structure reasons for his conclusion. Holmquist (1926) made a similar thin section examination of quartz lamellae. Although he found that many of them were bent he concluded that there was no clear evidence of translation. Other investigators—Judd, Milch, Fischer, and Johnsen—have considered the lamellae from various points of view without arriving at any unanimous conclusion.

The most recent work is by Hietanen (1938). She follows Mügge essentially in her analysis of lamellae-undulatory zones relations but gives a more detailed hypothesis as follows: Axial translation (bend-gliding) takes places parallel to the base (0001), the deformation probably producing elastic distortion (undulatory extinction) of the body of the crystal. The amount of translation is very limited, however, and the increasing resistance to plastic deformation forms minute openings in the glide surfaces. These form the visible evidence of the lamellae. Further deformation results in rupture and translation subparallel to the crystal axis so that the grain is divided into numerous submicroscopic "blocks" each capable of responding to deformation more or less independently of its neighbor. Slight rotation of these "blocks" results in change of orientation of the crystal as a whole without seriously disturbing the original position of the lamellae lines. This explains the anomalous orientation of the lamellae subparallel to (0001) as a relict orientation due to "change of front" (Frontwendung) of the submicroscopic ruptured parts. The amount of Frontwendung is not fixed, thus accounting for the irrational orientation of the lamellae. Rhombohedral translation is thought to take part in the general process.

This ingenious hypothesis does not consider the probabilities of quartz translation in terms of the crystal structure. There seems to be no greater likelihood of translation parallel to (0001) than on surfaces subparallel to it. The irrational lamellae support this conclusion. Assuming a translation hypothesis, it is only necessary to select a glide line, and study of the quartz crystal structure indicates that $[m{:}r]$ is the most probable one. The petrofabric evidence in support of this has already been presented.

CONCLUSION

It need scarcely be pointed out here that the mechanics of quartz orientation is inadequately understood, due largely to the ill-success of experimental work. This is directly related to the close-knit, immensely strong structure of the mineral which so far has defied almost all attempts to reproduce the deformation and recrystallization phenomena

found in nature. It was already stated on an earlier page that this lack of experimental data should not be a deterrent to hypotheses based on other evidence. The present paper contributes an additional fragment to the quartz problem through a correlation of field, petrofabric, and crystal structure data. The main results of the investigation are listed in the abstract. The merits of the translation hypothesis used to interpret these data cannot be accurately assessed at the present time.

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PETROGRAPHY AND ORIGIN OF THE REX CHERT

BY W. D. KELLER

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ABSTRACT

The Rex chert member of the Phosphoria formation of Permian age was studied petrographically from specimens collected across the stratigraphic sections at five localities in Idaho, Wyoming, and Utah. The Rex is a fine-grained quartzite with chert in its upper part in the Big Hole Mountains of Idaho. Southward toward the Utah-Idaho lines much of the clastic character is lost, calcareous chert predominates, and the member increases in thickness. It becomes thinner again farther south toward the type locality.

Microcrystalline quartz predominates in the chert, but some thin sections were studied in which three varieties of silica—clastic quartz grains, chalcedony, and microcrystalline quartz—all occurred in a single section. Petrographic evidence suggests that the chert of the Rex is primary with probably considerable diagenetic replacement of carbonate. It is inorganic in the main but contains some siliceous sponge spicules. Subsequent replacement is believed to be insignificant.

INTRODUCTION

The Phosphoria formation of Permian age, which includes the Rex chert member, has been extensively investigated by Mansfield (1927) and others, but the Rex chert has been studied petrographically to only

a limited extent. The writer offers a thin-section study of the most siliceous and resistant facies of the member and an interpretation of the petrographic data as to the origin of the chert.

The name Rex chert appeared for the first time in a table published by Richards and Mansfield (1912, p. 683), but they state that the selection of the name for the member was made by Gale (1910, p. 457-535) to whose work they refer. The type locality is Rex Peak, 5 miles east of Randolph, Utah, in the Crawford Mountains.

The Rex member occurs as a fairly distinct unit from the south side of the Uinta Mountains in Utah, northward probably into Alberta, Canada, and in an east-west direction from about longitude 108° to about 114°. According to Mansfield the maximum thickness seems to be developed in eastern Idaho and in some adjacent parts of western Wyoming.

Because of its pronounced resistance to weathering the Rex forms bold ridges, holds up steep and narrow canyon gateways, and provides a useful stratigraphic and physiographic marker for the economically valuable underlying phosphatic shale and phosphate rock. It possesses this outstanding resistance to erosion although in many places it is severely shattered into hand-sized and smaller fragments, chips, and blocks because of the brittleness of the rock and the severity of the folding and faulting common to the region of its outcrop. These pieces are usually held together in the bed by interlocking wedge action or by cementation with calcite.

At most outcrops the Rex is dark gray, black, or blue black but, where rich in carbonate or quartz silt, the color may range through lighter shades of gray. Lithologically it is usually compact and very fine-grained. Samples from some localities appear to the unaided eye as chert but are seen under a hand lens to be definitely quartzite. Other samples appear somewhat sugary or granular and cannot be classified with certainty without a microscope as either chert or quartzite. As the origin of chert and quartzite is so different, a record of the relative amounts of the two in the Rex should be interesting.

In order to determine the chert-quartzite relationship and other detailed characteristics of the Rex it was decided to study thin sections cut from hand specimens which were taken from several stratigraphic sections at every pronounced lithologic change in the stratigraphic section or at regular intervals where the lithology was more uniform. Sampling was restricted chiefly to the most resistant and cherty (or chert-like) part of the member because the main interest was in the chert rather than in the obviously noncherty, nonresistant shale or limestone portions. In consequence, localities were chosen where the chert was thickest and best exposed.

Samples were taken from five exposures in four rather widely separated localities. (See Figure 1.) The northernmost locality, No. 1, is in the South Fork of Mahogany Canyon (Schultz, 1918, Pl. 1) about 10 miles southwest of Driggs, Idaho, and a few miles west of the Teton Moun-

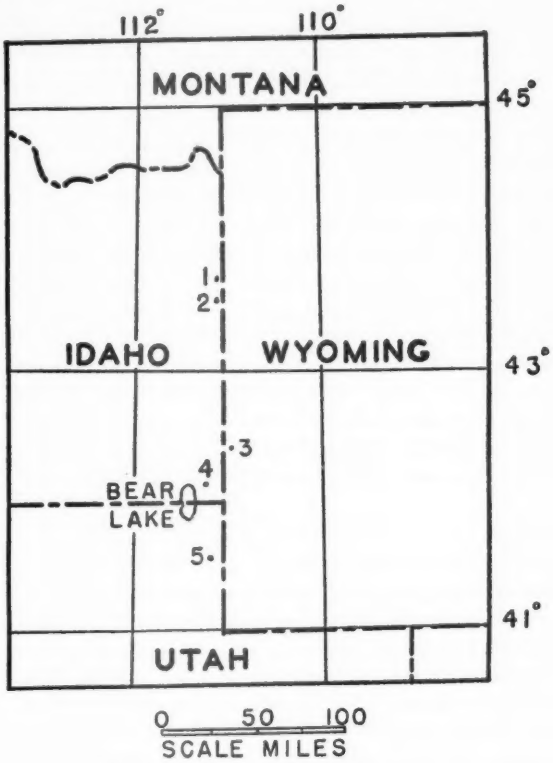

FIGURE 1.—*Index map of localities*

tains. Another suite of specimens was collected from Patterson Creek, locality No. 2, about 5 miles south of locality No. 1. Locality No. 3 is in Raymond Canyon in southwestern Lincoln County, Wyoming, which is about 85 miles south of locality No. 2. Locality No. 4, about 20 miles southwest of locality No. 3, is in the overturned beds at the northeastern corner of Bear Lake, Idaho. The last two sections are mapped by Mansfield (1927, Pl. 9). Samples were also taken from the

type section on Rex Peak, Utah, locality No. 5 (Gale and Richards, 1910), which is about 40 miles south of No. 4.

ACKNOWLEDGMENTS

The writer's interest in the Rex chert was aroused during the latter part of the summer of 1935 while he was a temporary assistant to James Steele Williams, of the United States Geological Survey, who was working in the western phosphate region. Very fortunately, G. R. Mansfield spent a part of his field season with Mr. Williams, and the stimulating discussions on the problems of the area, which arose from time to time, furthered a desire of the writer to study the Rex in greater detail.

Accordingly, a collection of material made at intervals across the stratigraphic section at Bear Lake Hot Springs was studied. More specimens from other localities were necessary, and during August 1938 additional material was collected from localities suggested by Mr. Mansfield and Mr. Williams. Part of the travelling expenses in 1938 and the preparation costs of thin sections were paid by a University of Missouri Research Grant. The thin sections of the 1935 collection were made with N. Y. A. help. Mr. Williams, Mr. Mansfield, and Mrs. W. A. Tarr have read the manuscript and have made valuable suggestions. For all the courtesies, advice, and monetary assistance extended him, the writer wishes to express his appreciation and thanks.

DETAILED DESCRIPTIONS OF SECTIONS

SOUTH FORK, MAHOGANY CREEK SECTION, IDAHO

On the north bank of South Fork of Mahogany Creek, near Driggs, Idaho, about one mile above its confluence with North Fork, the Rex chert is well exposed by stream erosion. Although its exact contacts with the adjacent formations are covered by talus, typical Woodside shale and phosphate rock (Phosphoria) are exposed about 100 feet up and down stream respectively. As both of these are weakly resistant to erosion there is little doubt that the chert is completely exposed.

On the outcrop the Rex is chiefly a gray, fine-grained quartzite with some patches or blotches of a more compact, cherty-appearing rock. The bedding planes are fairly distinct; thickness of beds ranges from 1 foot to 4 feet. Brecciation is severe, probably largely as a result of proximity to a large thrust fault. No abrupt changes in lithology were noticed in the entire thickness except within the 9-inch zone from which specimen M. C. 3 came, but samples were collected at intermediate positions for microscopical study which might reveal characteristics not visible to the unaided eye.

Sample No.	Detailed Descriptions	Feet above base of Rex
M. C. 5	Cherty quartzite, angular to subangular somewhat regenerated grains about 0.07 mm. in diameter cemented with microcrystalline quartz and sparse chalcedony. Cement occupies one-tenth to one-half of slide. Occasional carbonate or phosphate particle. Chert stained brownish gray by phosphatic material	21
M. C. 4	Sandy chert which has a greasy luster because of high content of clastic quartz (about 40 per cent), 0.05 to 0.15 mm. in diameter. Clastic grains are embedded in microcrystalline quartz and about 10 per cent fibrous chalcedony with an occasional phosphate pellet or carbonate rhomb	17
M. C. 3	Dark-gray, fossiliferous zone of calcite, chert, and phosphate rock. Brown, subrounded phosphate pellets, subangular to subrounded quartz grains about 0.15 mm. in diameter, and interstitial calcite are present in approximately equal proportions. Zone crowded with *Derbya (Plicatoderbya) magna* (Branson) identified by J. S. Williams, 1939	11
M. C. 2	Quartzite with a little carbonate and phosphate like M. C. 1	7
M. C. 1	Fine-grained quartzite (average diameter of grains 0.15 mm.) containing about 10 per cent scattered calcite and 5 per cent phosphate in very small, almost colorless grains or clusters. Original subangular quartz is now more angular. Little or no sutured texture is developed, and undulatory extinction is weak if present. Silicification, therefore, probably did not result from dynamic metamorphism but must have occurred during sedimentation or was supplemented by later work of ground water. No chalcedony or microcrystalline quartz were seen. Many quartz grains show fine, dark, parallel cracks which terminate at individual grain boundaries and show random orientation in the whole section. They were developed prior to deposition of the sand and confirm the clastic character of the lower part of the Rex	3

The lower half (10 feet) of the Rex chert on Mahogany Creek is almost entirely clastic. The upper half averages between one-third and one-half clastic quartz. For the entire section, roughly, one-fourth is nonclastic silica, and three-fourths clastic quartz.

A general note may be in order here about the nonclastic, very fine-grained (about .005-.008 mm. in diameter) silica, *i.e.*, microcrystalline quartz, which is abundant in M.C. 4 and in many specimens described later. The fine grains mutually interlock and show more or less undulatory extinction but do not have the fibrous character of chalcedony. Their index of refraction is close to that of canada balsam, and those favorably oriented show (using a 100X oil-immersion objective) an optically positive uniaxial interference figure. These optical properties indicate quartz. In previous studies of chert at the University of Missouri, the possibility was recognized that these fine grains showing undulatory extinction might be bundles of chalcedony fibers, but X-ray pictures made for Tarr (1936) gave a quartz pattern which confirms the optical results. In the writer's experience with cherts from a number of localities and formations microcrystalline quartz is their chief con-

stituent. Chalcedony, despite the statements in some textbooks, occupies a subordinate position.

PATTERSON CREEK SECTION

On Patterson Creek, about 5 miles west and 2 miles north of Victor, Idaho, the Rex chert crops out in a bluff from which seven samples were taken. As at the Mahogany Creek locality, the hand specimens were collected at changes in lithology and at regular intervals where the rock showed little variation. Very little difference was apparent between the lithology of this section and that previously described from Mahogany Creek. The beds here dip 20° S. 35° W.

Sample No.	Detailed Descriptions	Feet above base of Rex
P. C. 7	Sandy chert, shown in Figure 1 of Plate 1. Composed chiefly of a fine-grained microcrystalline quartz matrix in which are embedded subrounded clastic quartz grains and chalcedony aggregates with concave border indentations. Hence, three varieties of silica are in one thin section	20
P. C. 6	Sandy chert. Light-tan microcrystalline quartz matrix containing 15 or 20 per cent angular to subangular clastic quartz grains scattered uniformly throughout. On a few grains microcrystalline quartz has grown radially from the surface	15.5
P. C. 5	Phosphatic quartzite. About 20 per cent phosphate occurs in a gray quartzite made of cracked, subangular, well-interlocked quartz grains averaging about 0.2 mm. in diameter. The phosphate occurs in tan grains about the same size as those of the quartz and also in little clusters of small clear grains	13
P. C. 4	Quartzite like P. C. 5 except that the phosphate content is about 10 per cent, and microcrystalline quartz aggregates about the size of the sand grains compose approximately 5 per cent of the section	10
P. C. 3	Quartzite like P. C. 4 without microcrystalline aggregates	7
P. C. 2	Quartzite similar to P. C. 3	2
P. C. 1	Quartzite similar to P. C. 3	4

At Patterson Creek, as at the Mahogany Creek locality, the Rex is a quartzite in its lower half. The upper half of both localities, however, contains more microcrystalline quartz (chert) with less clastic material than the lower half.

RAYMOND CANYON SECTION

A spectacular, almost vertical, high wall of the Rex forms the narrow gateway of Raymond Canyon where it enters the broad valley of Thomas Fork of Bear River. The outcrop is ideal for study and for sampling the entire thickness (70 feet) of its exposure. Less clastic material is present here than farther north, but much carbonate (chiefly calcite) occurs in the member. In certain beds the carbonate predominates; in others, the rock is chiefly chert. Usually, one is distributed within the other as nodules, blotches, or discontinuous and irregular bands

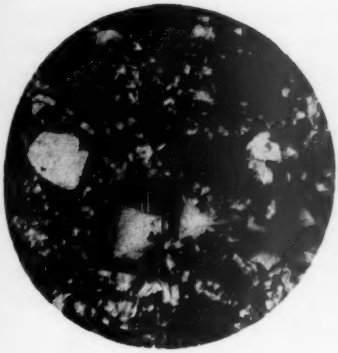

FIGURE 1. CLASTIC QUARTZ (LARGE WHITE AND BLACK AREAS), CHALCEDONY (FIBROUS CROSSES), MICROCRYSTALLINE QUARTZ (MOTTLED)
Crossed nicols. x 65.

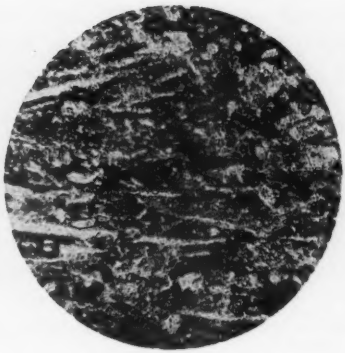

FIGURE 2. SILICEOUS SPONGE SPICULES
Transmitted light. x 65.

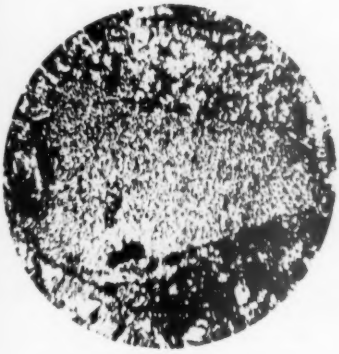

FIGURE 3. CHERT MICROBRECCIA
Crossed nicols. x 22.

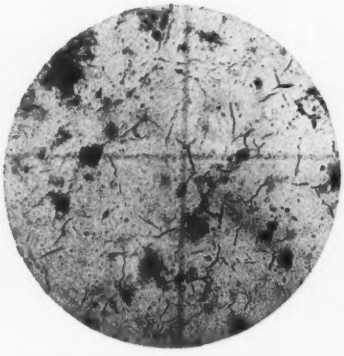

FIGURE 4. TENSION, DESICCATION (?) CRACKS IN CHERT
Transmitted light. x 65.

PHOTOMICROGRAPHS OF REX CHERT

rather than as finely disseminated material. On the outcrop the carbonate is commonly gray, recessed by weathering, or surface pitted, whereas the chert is usually black or very dark gray and uncorroded by weathering.

A number of beds or zones each of which appears uniform but different from those adjacent are easily recognized. Representative samples were taken of each of these distinctive zones, usually from both the gray carbonate and the black chert, but where the zones were quite thick more than one pair was taken. Thicknesses of the zones in descending order are given with the detailed descriptions of the chert and carbonate samples.

Sample No.	Detailed Descriptions	Zone thickness in feet
	One massive black chert bed carrying bulbous nodules and irregularly wavy strips of gray limestone which makes up about one-fourth of the bed. The bulbous forms, which are up to 15 inches in cross section, show long axes predominantly at right angles to the direction of stratification.	
R. C. 36	Chert is about four-fifths microcrystalline quartz mottled tan. Carbonate in clay-sized particles, irregular small aggregates, and rhombs fills the rest of the section. Section is moderately fossiliferous with sponge spicules.	
R. C. 37	The limestone is fine-grained, anhedral, and contains about 15 per cent of fine nonclastic quartz distributed fairly uniformly throughout the section	10
	Massive bed of black chert containing roughly one-fourth limestone distributed in patches one to six inches thick with their long dimensions parallel to the bedding.	
R. C. 34, 32	Chert is similar to that in R. C. 36. Upper part of zone is highly fossiliferous, half of the microcrystalline quartz being sponge spicules most of which show no central canal.	
R. C. 35	Fine- to medium-grained limestone from near R. C. 34. It contains irregular spots of very coarse-grained chert which show transitional contact with the limestone and the curved "caries" shape border common to replacement texture. This coarse chert differs from the usual microcrystalline quartz predominating in most of the rock.	
R. C. 33	Fine-grained carbonate with about 10 per cent each microcrystalline quartz and phosphate	14
R. C. 31	Black chert layers, 2 to 6 inches thick, separated by thin gray partings. Thin section is 60 to 70 per cent microcrystalline quartz, the remainder being very fine-grained carbonate more or less stained with dark phosphatic material, and a few carbonate rhombs. Thin laminations are shown by alternating layers of clear, silica spicules and a dark material. (*See* Figure 2 of Plate 1.)	6
R. C. 30	Massive black chert beds, one to two feet thick, with small lines and irregular partings of gray limestone. The chert is 70 to 80 per cent microcrystalline quartz with the remainder extremely fine "dust" and a few small carbonate rhombs which are concentrated with dark phosphatic material between siliceous sponge spicules. The canal type of spicule predominates but the solid type is also present.	
	A small part of the section is carbonate containing sponge spicules without canal structure	5

Sample No.		Zone thickness in feet
	Black nodules, 2 to 6 inches thick, set in a gray matrix, in a ratio of about three to two, show wide, discontinuous bedding in this zone.	
R. C. 29	The gray matrix is a fine-grained limestone containing about 25 per cent microcrystalline quartz.	
R. C. 28	The black nodular rock is fine-grained siliceous limestone in which the microcrystalline quartz (30 to 40 per cent) occupies chiefly broken sponge spicules	11.5
	Poorly defined, ellipsoidal nodules of gray limestone, about 2 to 8 inches long and from 1 to 3 inches thick, are set in a matrix of black chert; ratio of limestone to chert is about one to four.	
R. C. 27	The limestone is medium-grained, fossiliferous with a few carbonate spicules which show no canal structure, and carries about 5 per cent microcrystalline quartz and a small amount of fine quartz sand.	
R. C. 26	The chert is composed almost entirely of silica with sparse carbonate. Sponge spicules, most of which have interior canals, make up approximately one-third of the section. They average 0.15 mm. in diameter and are largely chalcedony although microcrystalline quartz may occupy part or even all of an area of spicules	6
R. C. 25	Chert and limestone nodules, 2 to 4 inches in cross section and about equal in number, give a spotted appearance to this zone. A thin section cut across a chert-limestone contact shows the contact to be sharp. The chert is essentially all microcrystalline quartz with only a few carbonate rhombs, and the limestone is chiefly medium-grained calcite with small amounts of fine microcrystalline quartz scattered through it	7
R. C. 23, 24	This zone appeared in the field to be about half black chert and half gray limestone, but the thin sections showed both parts to be limestone with dolomite rhombs set in anhedral calcite, and microcrystalline quartz varying from about 40 to 10 per cent. A few sponge spicules were present	2.5
R. C. 22	Highly brecciated black chert makes up this zone. It is composed chiefly of fine, microcrystalline quartz but contains 5 to 10 per cent carbonate rhombs about 0.07 mm. in cross section, very fine carbonate "dust," and occasional lumps of dark material	2
	This zone at the base of the Rex is a grayish, pitted, highly fractured limestone blotched by patches of black chert 6 to 14 inches across which retain sharp angular edges and plane faces on weathered surfaces. Chert makes up one-fourth to one-fifth of the zone.	
R. C. 21	The limestone is made of irregular, mostly fine, interlocking carbonate grains and about 5 per cent quartz silt particles 0.05 to 0.1 mm. in diameter, many of which show a little secondary enlargement. A small part of a trepostomatous bryozoan and a few sponge spicules without internal canals are present.	
R. C. 21A	The chert is chiefly microcrystalline quartz, about 10 per cent of fine dolomite rhombs, a little quartz silt, and a few sponge spicules. The microcrystalline quartz is spotted by gray to brownish phosphatic material	6

At Raymond Canyon the Rex is almost entirely nonclastic. Limestone, and chert which is commonly calcareous, constitute the lower fourth of the

member. Chert with less carbonate, either as an admixed impurity or in megascopic occurrence, comprises the upper three-fourths. Fossil sponge spicules make up an appreciable part of the member at this locality. The presence of abundant nodules, either of limestone in chert, or of chert in limestone, is notable.

BEAR LAKE HOT SPRINGS SECTION

A good section of the Rex is exposed at the northeastern corner of Bear Lake in a valley due east of the Hot Springs bath house where the beds are overturned, having an apparent dip of 51° N. 86° W. Representative samples were easily obtained from the entire thickness of the resistant portion, but the contact of the upper flinty shale facies of the Rex with the overlying Woodside shale (Triassic) was not determined exactly because of cover by talus.

Specimens were collected across the 300 feet of exposure in this valley, but only the 16 samples which came from the resistant part of the Rex, 141 feet, 8 inches, are described in detail.

A section across the beds above the resistant Rex, which have been referred to by Mansfield (1935) as "probably representing the flinty shale facies of the Rex," is given herewith:

FLINTY SHALE OF THE REX AT BEAR LAKE HOT SPRINGS

	Thickness of zones in upper, nonresistant part of Rex, Feet
Shale poorly exposed, contact (?) of Rex chert and Woodside shale 300 feet above base of Rex	38
Yellow shale and siltstone	17
Unexposed. Lithology of talus is like that of the bed above	36
Thin, platy, yellow shale	26
Fine, gray quartzite	9
Platy, dark-gray shale	3
Gray quartzite and gray chert	17
Siltstone and shale	3
Chert and quartzite (dark gray)	8
Dark-gray siltstone	7

The flinty shale section is chiefly clastic although much of the quartzite is so compact and fine-grained that it superficially resembles the true chert in the member. In fact, the resemblance on the outcrop to chert was so close that it was sampled with the thought that it was simply a less resistant chert, but when studied under the microscope it was seen to be actually a quartzite composed either of fine sand or silt with additional clay, phosphate, and fine, usually colorless mica flakes.

The resistant lower cherty part of the Rex is described in detail as follows:

Sample No.		Zone thickness Feet	Inches
H. S. 17	A massive bed of translucent bluish-black chert, 5 feet thick, marks the top of the cherty Rex at Bear Lake Hot Springs. It is composed of unstained fine microcrystalline quartz and up to 20 per cent carbonate rhombs averaging about 0.07 mm. in length. Shrinkage cracks like those described in H. S. 21 are abundant	5	
H. S. 18	Phosphatic silty chert in which the microcrystalline quartz (0.02 to 0.03 mm. in diameter) carries about 15 per cent clastic quartz grains (0.06 to 0.08 mm.) and a small amount of non-opaque phosphate mineral aggregates and dark phosphate stain. The clastic quartz may show regeneration by ordinary quartz or by a layer of radial microcrystalline quartz which is in as close optical continuity with the center as can be recognized in the undulatory extinction	2	6
	This thick uniform zone is most striking in the field where it stands out from the mountain side as a high, prominent, black, shiny wall because of its superior resistance to weathering. On the outcrop it is devoid of any decomposition product—suggesting a high content of relatively pure compact silica. On close view it is seen to be composed of irregular, gnarly, subtranslucent, blue-black beds of chert varying in thickness from 4 to 18 inches.		
H. S. 19	Top of the zone. This thin section is similar to H. S. 21, with a few sponge spicules.		
H. S. 20	Taken 9 feet below H. S. 19. Similar to H. S. 21.		
H. S. 21	Taken 10 feet below H. S. 20. The thin section, shown in Figure 4 of Plate 1, shows the chert to be almost pure silica in extremely fine grains of microcrystalline quartz, 0.007 to 0.020 mm. in diameter, approaching cryptocrystallinity. Radiating tension (shrinkage?) cracks have been developed possibly through desiccation of a silica gel which gave rise to chert. The specimen is nonfossiliferous.		
H. S. 22	Taken 15 feet below H. S. 21 which it resembles except for having a little coarser quartz. The purity of the chert is noteworthy.		
H. S. 23	Taken 15 feet below H. S. 22. This is very fine microcrystalline quartz (0.015 to 0.030 mm. in diameter), slightly fossiliferous with sponge spicules, and pure silica except for a very little light greenish phosphate stain.		
H. S. 24	Taken 15 feet below H. S. 23. The section is almost pure microcrystalline quartz with a few coarse grains about 0.09 mm. in diameter. A wavy dark line, like a cross section of a micro-sized stylolite, cuts across part of the section.		
H. S. 25	Taken at the base of the Rex, 11 feet below H. S. 24. This section is also almost pure fine microcrystalline and a little chalcedony. A few very small carbonate rhombs and spots of phosphatic stain are disseminated through it	85	
	This zone is composed of irregular chert beds 4 to 6 inches thick which resist weathering weakly.		

Sample No.		Zone thickness Feet	Inches
H. S. 26	The thin section is chiefly microcrystalline quartz stained with dark-brown calcareous phosphate. Sponge spicules make up about half of the section. They show central canals usually, are about 0.2 mm. in length, and vary from 0.04 to 0.08 in diameter. A few large subangular clastic (?) quartz grains are also scattered through the section	12	
H. S. 27	Dark-gray, dull, slightly granular chert which weathers readily. It is similar to H. S. 26 in thin section except that a few large, well-preserved carbonate sponge spicules with central canals are present	3	
H. S. 28	A zone of broken chert beds 4 to 6 inches thick which weather to small, angular, chippy fragments. It is similar to H. S. 26 in thin section except that no clastic quartz grains are present	21	
	A sandy, dark-gray shale which contains large lenticular black nodules which, because of their elongation, appear to form a discontinuous bed.		
H. S. 29	A thin section of a nodule shows it to be a slightly siliceous, phosphatic limestone stained dark brown. The silica, 10 per cent of the section, is in fine quartz grains scattered through the section and in microcrystalline quartz sponge spicules without central canals	2	6
	This zone is a 3-inch chert layer overlying 4 inches of shale.		
H. S. 30	The chert is about half microcrystalline quartz, and half chalcedony in sponge spicules ranging from 0.1 to 0.3 mm. in diameter. Brownish phosphatic material stains the areas between the fossils.		
	A few relatively large (0.5 to 0.8 mm. in diameter) angular fragments of fine microcrystalline quartz aggregates are embedded in the coarser microcrystalline quartz and chalcedony of the section. In other words, the sample is in part a breccia. (See Figure 3 of Plate 1.) The earlier fragmented chert was very fine-grained; the later chert matrix is coarser-grained and fossiliferous. The writer interprets this as an intraformational microbreccia whose significance lies chiefly in the evidence that chert available for fragmentation was developed very early in Rex history. It offers good evidence for a primary origin for at least part of the Rex chert.		
	Small hexagonal euhedrons of a phosphate mineral, probably francolite, are also visible in the thin section. Fragments of the crushed sample, concentrated in bromoform, had indices of refraction varying from 1.618 to 1.628, low birefringence, and showed very weak negative interference figures. These data suggest, possibly, francolite		7
	One thick bed of black chert, massive except for fine, irregular, short, discontinuous partings 6 to 10 inches apart, and spotty patches of limestone make up this zone.		
H. S. 31	The chert is about 80 per cent microcrystalline quartz and 20 per cent carbonate in irregular grains about 0.08 mm. in diameter and in a few small rhombs. The limestone, stained light brown, is composed of both calcite and dolomite and about 10 per cent subangular quartz grains	9	

The resistant Rex chert at Bear Lake is dominantly chert and includes but little clastic material or carbonate, differing in this respect from the

other sections described. Sponge spicules are present in several zones, but the most prominent and thickest zone of chert (85 feet thick) contains few fossils. The chert in this prominent zone is rather pure, very fine-grained, compact, and its quantity, if the zone extends over a reasonable area, must run up to immense figures. The occurrence of a chert microbreccia in another zone shows the presence of at least two generations of chert.

REX PEAK SECTION

Rex Peak, the type locality of the Rex chert member, is in the Crawford Mountains about 5 miles east of Randolph, Utah. The structure here is complicated by strong thrust faulting and severe folding, in which compact, brittle chert is adjacent to softer, less competent shales. The structure may, therefore, change greatly within short distances. Across the outcrop on the peak, residual gravel has covered much of the member, and only intermittently do more resistant beds protrude, and then perhaps for only a few inches. The dip cannot be taken with certainty in the lower part of the formation because of poor exposures. Because of all these difficulties, obviously an accurate measure of the thickness is not possible. At the top of the member, a dip of 17° E. was measured. If this dip prevailed across the peak, about 75 feet of Rex would be present. However, Gale and Richards (1910, p. 476) give a thickness "from 125 to 200 feet at various places in the Crawford Mountains," but erratic or incomplete thickness may be expected because of the faulting present here. James Steele Williams (personal communication, 1935) has advised that a transecting fault in the type locality necessitates extreme care in accurately measuring the Rex.

Obviously, considerable time would be needed to make as detailed a section here as was done at the other localities, and because the writer could not spend the necessary time in the region measured sampling was not done. Instead, it was decided to collect from the base of the Rex (which could be fairly accurately located) and at each succeeding bed which protruded through the soil, without attempting to make a strictly quantitative description of the section. In all, 17 samples were taken within moderately regular intervals, and these are described in descending order to show in a qualitative way what makes up the Rex across its type section.

Detailed Descriptions

Sample R. P. 17, taken close to the top of the Rex, is a large specimen made up of irregular rounded bodies of chert and carbonate molded or deposited together so that the constituent nodules indent each other mutually.

A thin section cut across the contact of the two shows the carbonate to be a fine-grained dolomitic limestone with probably 5 per cent quartz silt scattered through it, and the chert to be fossiliferous (Bryozoa) microcrystalline quartz and chalcedony.

The fossiliferous structure can be easily recognized in the chert up to the contact with the dolomite but it is traced only questionably out into the carbonate, probably because of the recrystallization of the limestone. It appears, therefore, that only part of the bryozoan was replaced before recrystallization of the limestone occurred. The time of the limestone recrystallization cannot be dated positively, but, if this took place during diagenesis, the silicification of the bryozoan antedated this stage in diagenesis.

R. P. 16 is microcrystalline quartz with perhaps 5 per cent regenerated fine sand grains.

R. P. 15 is chiefly microcrystalline quartz, with some chalcedony in fine radial segments and bundles, and occasional regenerated fine sand grains.

R. P. 14 in hand specimen is light-gray chert which weathers to a finely porous surface studded with resistant grains or aggregates. In thin section it is seen to be chiefly microcrystalline with 10 to 20 per cent strongly regenerated fine sand grains whose diameters have been almost doubled. Irregular aggregates of carbonate, about 0.1 mm. in diameter and amounting to about 10 per cent, and a few rhombs are scattered through the section. A very little phosphate is also present.

R. P. 13 is similar to R. P. 14 without carbonate rhombs.

R. P. 12 is composed of over 95 per cent silica about equally divided between fine-grained chalcedony and microcrystalline quartz.

R. P. 11, 10, 9, 8, are very fine microcrystalline quartz with 30 to 40 per cent carbonate dust and rhombs.

R. P. 7 is dark-gray chert with a slightly brownish hue. In thin section it is found to be microcrystalline quartz with 10 to 20 per cent carbonate in very fine grains and a few rhombs. A considerable part of the thin section is a portion of cryptostomatous bryozoan, identified by Professor R. E. Peck, which has been replaced by microcrystalline quartz and chalcedony.

R. P. 6 is about 60 per cent very fine microcrystalline quartz, the remainder being carbonate in dust and fine-grained rhombs, and an occasional grain of phosphate.

R. P. 5, 4 are similar to R. P. 7.

R. P. 3 is a black, fine-grained dolomite, rhombs varying from 0.02 to 0.07 mm. in cross section, and about 5 per cent clastic quartz in grains 0.03 mm. or less in diameter. A little chalcedony is also present.

R. P. 2 is a dark-gray, compact, badly shattered chert made of tan-mottled microcrystalline quartz, with 10 to 20 per cent carbonate in very fine grains and a few shattered rhombs.

R. P. 1 is similar to R. P. 3.

The part of the Rex chert most resistant to erosion at its type locality is a blue-black chert which is composed chiefly of microcrystalline quartz with some chalcedony. Carbonate is abundant in the rocks, and clastic quartz is present in lesser amounts. A few bryozoan fossils were also seen.

CHERT SAMPLE FROM PHOSPHORIA FORMATION

A single chert nodule was collected from the Phosphoria formation at Bull Lake on the northeastern side of the Wind River Mountains in Wyoming. Although the lithology of the Phosphoria here is entirely different from that in Idaho-Utah sections, the equivalence has been demonstrated by paleontologic evidence. No resistant chert zone is found in the Wind River Mountains corresponding to the one described

from the region to the west, but cherty beds and siliceous portions of the Phosphoria are abundant in the Wind River section.

A thin section of the nodule collected at Bull Lake shows very fine microcrystalline quartz, almost cryptocrystalline, with a few scattered quartz grains of about silt size. Carbonate in small amount and in fine grains is also seen. A few mica flakes lie in parallel orientation, and some phosphate stains the chert locally.

GENERAL SUMMARY

The Rex chert in its distribution along longitude 111° 30′ varies considerably in the most resistant part of the member. Near Driggs and Victor, Idaho, the resistant part, which is relatively thin (21 feet thick), varies from a quartzite, definitely clastic at the base, to sandy chert and chert in its upper part.

Ninety miles south, in Raymond Canyon, Wyoming, the resistant Rex is 70 feet thick, has lost most of its clastic character, but contains abundant carbonate in addition to the silica. Nodular structures in both carbonate and chert, and fossil sponge spicules in the chert, are common.

At Bear Lake Hot Springs, only 20 miles south of Raymond Canyon, the thickness of the same resistant part has increased to about 140 feet. Outstanding here is a zone 85 feet thick standing in a tilted wall of resistant, subtranslucent, bluish-black chert, which is relatively pure silica.

At Rex Peak, the type locality of the Rex member, about 40 miles south of the Bear Lake locality, more carbonate was seen mixed with the chert than at Bear Lake.

The chert of the Rex member, taken as a whole, is composed chiefly of microcrystalline quartz, a fine to very fine-grained silica which commonly shows undulatory extinction but otherwise possesses the optical properties of ordinary coarse-grained quartz. The microcrystalline quartz may constitute almost the only material present, it may occupy sponge spicules, or it may serve as a matrix in which clastic quartz grains, carbonate dust and rhombs, or phosphatic material are embedded. In calcareous or dolomitic portions the microcrystalline quartz particles are commonly disseminated throughout the limestones. Chalcedony, which is much less abundant than microcrystalline quartz, occupies sponge spicules or is scattered in radial sectors or aggregates in the microcrystalline quartz. Within the writer's experience, after studying thin sections of many cherts, the total of microcrystalline quartz in them far exceeds the amount of chalcedony. Some textbooks leave an erroneous impression that chalcedony is the essential mineral of chert.

The limestone in the Rex is either calcareous or dolomitic. The ratio was not determined. It is suggested from limited tests that when embedded in chert the euhedral rhombs are dolomite and the anhedral grains are calcite.

Clastic quartz grains have been recognized in the chert by the cracks developed in them before deposition in the chert, by their coated boundary surfaces, and by their inclusions. Regeneration is also common.

Tan to dark-brown phosphatic material has stained the chert. The phosphate appears concentrated with carbonate but is scantily associated with silica in a number of thin sections. Some transparent hexagon-shaped grains of phosphate mineral are tentatively identified as francolite. A few mica flakes were found.

Because the phosphate rock in the Phosphoria formation is blue-black and because of its close association with the chert, it seems reasonable that organic phosphate is responsible for the dark color of the Rex chert.

PETROGRAPHIC EVIDENCE ON THE ORIGIN OF THE REX CHERT

The origin of chert has been a topic of interest for many years, and a subject of active discussion between investigators who saw in their field and laboratory studies evidence for its origin by replacement and others who interpreted the evidence they saw as proof of a primary origin.

The writer was interested in what evidence might be obtained from a petrographic study which would bear on the origin of one of these thick and widespread formations and therefore offers the following with interpretations as applicable to the origin of the Rex chert.

(1) In the Bear Lake Hot Springs section a zone 85 feet thick is composed almost entirely of chert which is remarkably free from impurities, even in microscopic particles. The constituent mineral is microcrystalline quartz. Carbonate is notable in beds above and below.

The writer believes it is unreasonable to account for this great thickness of relatively pure chert under a late replacement theory because of the difficulties encountered in attempting to explain the mechanism by which such thorough and large-scale replacement could occur.

First, if the silica-bearing, replacing solutions moved down or up in sufficient quantity to replace 85 feet of rock they would probably have completely replaced all the carbonate en route. Obviously such thorough replacement was not done because the beds above and below the 85-foot zone contain carbonate mixed with the chert.

Second, if the solutions moved laterally instead of vertically they would have left along the margins of their channels irregular surfaces, or zones gradational from chert to limestone or other unreplaced rock, as are usually developed in replaced areas. No such contact zones were

seen. Further, if the solutions moved laterally, why and how were they able to do so over an area as large as that now occupied by chert? What was the source of the tremendous amount of silica now making up the Rex which is reconcilable with lateral transportation? Less serious but also disturbing problems are connected with the avenues of removal and the disposal of the tremendous quantity of carbonate replaced. The writer finds no satisfactory answers to these questions. Also, where the Rex contains both limestone and chert it is spotted light and dark, but in the purer 85-foot-thick zone the chert is rather uniformly dark. Had the 85-foot-zone undergone late replacement it seems probable that coloration would not be so uniform.

An alternative theory of origin of the chert as a primary deposit of silica on the sea floor supplemented by an indeterminate amount of diagenetic replacement appears more reasonable. The writer sees no criteria by which the extent of diagenetic replacement may be evaluated. Little or no replacement is necessarily indicated in the development of the 85-foot relatively pure chert zone, but in others where nodules of chert and carbonate are intermixed diagenetic replacement may have been considerable. Where fine grains of microcrystalline quartz and carbonate are mutually disseminated they may have accumulated as direct ocean precipitates or through partial replacement of carbonate before consolidation.

The source of silica for such a deposit does not present a difficult problem. Weathering land surfaces, even those high above sea level today, contribute an enormous amount of silica which is carried to the sea, as is shown by the average analysis of present-day river waters wherein silica is the second most abundant substance carried to the sea in solution. In Rex time it is probable that similar quantities of silica were carried into the sea to be deposited. Also, silica may have been given off in volcanic emanations, as suggested by Mansfield. The problem of disposition of all the silica brought in is solved plausibly by its deposition on the sea floor. Petrographic evidence in itself offers no clue as to whether the source of silica was volcanic action or weathering land.

Briefly, from a quantitative standpoint, a primary origin will readily account for the great quantity of widespread bedded Rex chert, whereas under an epigenetic replacement origin very serious, if not insurmountable, difficulties are encountered.

(2) The abundance, widespread occurrence, and fairly uniform distribution, as in the Rex, of either fine-grained carbonate in a dominantly siliceous zone or of fine-grained microcrystalline quartz in a dominantly carbonate zone is interpreted as being either a primary deposit in which silica or carbonate was precipitated in excess of the other or one in which

diagenetic replacement was effective. Secondary replacement, in contrast, is usually associated with irregularity of contact and with gradational zones. For instance, if one of the minerals occurred in fine particles as within a matrix of the other as "outliers" of a solid advancing wall (or perhaps continuous except for "islands" of the "other" mineral), the scattered distribution might be evidence of secondary replacement. The notable absence of such definitely recognizable replacement textures, even though it is negative evidence, becomes weighty because of the many specimens and thin sections studied and weakens the idea of late replacement.

The writer visualizes the deposition of a mixture of limestone and chert as an essentially simultaneous deposition from sea water which is saturated with both, supplemented by diagenetic replacement. No criterion occurs to the writer which may be applicable to a quantitative estimation of the amount of this replacement.

(3) At the Raymond Canyon locality, particularly, nodular or bulbous forms of chert were enclosed in or associated with limestone, and similar forms of limestone were set in or associated with chert. The nodular form of chert has been taken in certain cases as a criterion for its origin by late replacement, but if the criterion is applied broadly here then the limestone nodules lying in a dominantly siliceous bed would indicate their origin by late replacement of chert by calcium carbonate, an innovation indeed. The mutual inclusion of nodules of chert and limestone, each within the other, seems to be evidence against a unilateral late replacement, *i.e.*, origin of chert by late replacement of limestone.

On the other hand, the theories commonly entertained for the development of nodular limestones are that the quantity of calcium carbonate was insufficient, or the environment of deposition was unfavorable for the formation of a continuous bed, or that nodules developed during diagenesis. These same theories are preferably applicable to explain the more or less simultaneous deposition of nodular chert and limestone.

Tensional cracks forming a pattern like that developed by drying muds were seen in thin sections from samples collected in the thick chert zone at Bear Lake Hot Springs. (*See* Figure 4 of Plate 1.) Tarr, 1926 (p. 42) wrote, "Cox, Dean, and Gottschalk (1916) give 'shrinkage cracks' as evidence that chert was originally a gel. They found such shrinkage cracks in chert from the Joplin, Missouri, area." Tarr interpreted a silica gel origin of chert as primary.

In connection with the desiccation of silica gel and its change to crystalline quartz, a bit of speculation may be made on the cause of the undulatory extinction seen in the microcrystalline quartz of chert. The writer is unaware of any published explanation for the undulatory ex-

tinction of the quartz in chert, but in the case of metamorphic quartzites, where undulatory extinction occurs commonly, the explanation is given that the quartz is under strain, the space lattice is deformed, the optical indicatrix is distorted, and, therefore, extinction is not sharp. Following the same line of reasoning, perhaps the quartz in chert is under strain. The strain may have developed as the silica passed from the colloidal state in which it was deposited to a crystalline state. The strain would be within the space lattice and might develop undulatory extinction in the mineral.

Possible support for this idea lies in the additional evidence of the existence of strain in chert on a megascopic scale. It is a well-known fact that when a piece of chert is broken the two pieces usually will not fit together exactly when pressed together again. As early as 1898, Shepard (1898, p. 112) wrote,

> "When masses of this chert are broken into two pices, the parts, when replaced, do not exactly fit, indicating a molecular tension or strain."

The hypothesis that strain is developed in each crystal of microcrystalline quartz during the transition from colloidal to crystalline silica, that undulatory extinction is thereby occasioned, and that the summation of the intracrystal strains accounts for the strain in a nodule or large mass seems not too far fetched to be considered. However interesting the concept may be, its bearing on the origin of chert is probably not diagnostic because either primary chert or that orginating by replacement may have been carried in a colloidal solution and may have developed strain in the same way.

(4) In thin sections which show tan-to-brown phosphatic stain distributed within microcrystalline quartz and carbonate, the stain commonly appears to be concentrated in the carbonate-rich portions. It seems to the writer that, if the microcrystalline quartz originated as a late replacement of carbonate, the stain would be distributed evenly within both silica and carbonate. Selective replacement might avoid phosphate-rich areas, but if such selective replacement did occur it has been notably ineffective because an appreciable amount of stain is surrounded by silica. It seems more logical that the silica, calcium (and magnesium in dolomite) carbonate, and calcium phosphate were deposited essentially at one time and that the present close association and concentration of phosphate with carbonate is due to the effect of a common ion, calcium.

(5) The presence of very small angular fragments of cryptocrystalline quartz (chert) in a matrix of microcrystalline quartz (chert) in sample H. S. 21, forming a chert microbreccia, indicates an early origin for the

chert in the fragments, certainly not replacement at the same time as the matrix, if the matrix ever was replaced. The writer believes, on the other hand, that the breccia is intraformational and the fragments are part of an earlier, already consolidated chert, now set in a younger, also primary chert.

(6) Some secondary or late replacement of limestone probably did occur because the writer found in the section from sample R. C. 35 one example of gradational replacement texture. However, from a statistical basis the paucity of typical replacement texture as contrasted with the many thin sections examined supports the belief that by far the greater part of the Rex chert was formed as a primary deposit supplemented by a considerable but quantitatively indeterminate amount of diagenetic replacement. In this connection Mansfield (1927, p. 371) found that "little evidence has been found to support the ideas of replacement of diffusion."

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